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Sea-Level Changes in the Geological Record

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Abstract – No direct dependence of transgressions and regressions on sea-level position exists. Both regional tectonic movements, resulting in relative sea-level changes, and the supply of terrigenous clastic material can significantly complicate the global eustasy. The latter, however, left a clear impact on the geological record and, foremost, on the structure of sedimentary complexes (cyclothems). Eustatic fluctuations of the World Ocean level were of great importance in the formation of sedimentary basins.

E. Suess was the first to introduce the term *eustatic fluctuations* in 1906 for the global sea-level rises and falls that, according to him, equally with regional tectonic movements, could have conditioned the cyclothem formation. He suggested several approaches to their study, in particular, that based on tracing the ancient sea boundaries and analyzing sedimentary formations. The Suess ideas, however, did not win many followers, because at the beginning of this century, no objective criteria existed for the discrimination of eustatic "traces" from those caused by regional tectonic movements.

However, geological investigations collected more and more evidence that eustatic fluctuations greatly influenced sedimentary rock formation in the past. A number of interesting publications appeared in the second half of this century that demonstrated the rather frequent reflections of past sea-level changes on the geological record and, foremost, in the cyclic character of sedimentation. Global transgressions and regressions are thought to be related to these changes. At the same time, the direct correlation of these phenomena encountered certain difficulties, causing different paleogeographic interpretations. Well known, for instance, is the supposition that transgressions and regressions were diachronous (Yanshin, 1973).

It is scarcely possible to overestimate the importance of the studies, carried out by P. Vail and others in the 1970s and 1980s, in defining the role of eustatic fluctuations in the World Ocean history (Seismic Stratigraphy ..., 1977; Vail, 1987; Vail *et al.*, 1987). These studies showed for the first time the complex relationships of sea-level changes, transgressions, regressions, and sedimentation cycles. They also demonstrated that underestimation of these relationships may result in wrong paleogeographic reconstructions. According to recent data, eustasy conditioned by changes in water volume or World Ocean capacity is the main process that caused the sea-level changes. The second reason for the relative sea-level changes relates to the regional tectonic movements that affected ancient depositional basins or their parts and resulted in changes in bottom hypsography. Finally, the third reason for the relative drops in sea level and the migration of the shorelines is the supplying of terrigenous clastic material into the basin.

METHODS OF DEFINING RELATIVE SEA-LEVEL CHANGES

Reconstructions of sea-level position can actually be performed based on studies of the paleoshelves, particularly in zones of their conjunction with the land. The equilibrium between the accumulation and bottom erosion that exists within the shelf is distorted because of tectonic movements or eustatic sea-level changes. The subsidence of the bottom or sea-level rises are accompanied by active depositional processes. The uplifts of the shelf bottom or the sea-level drops stimulate the development of bottom erosion because of the wave activity. Some distortion of equilibrium and local accumulation may also be caused by the supply of a large amount of terrigenous clastic material.

Cratons, where subsequent tectonic movements almost do not distort the original sedimentary sequences, are of great importance for the paleogeographic reconstructions. The principles of facies analysis were assumed as the basis for the compilation of paleogeographic maps of continents (Schuchert, 1957; Metody ..., 1963; Atlas ..., 1967 - 1969; Atlas ..., 1992; Atlas ..., 1985). These compilations allowed defining the epochs of global transgressions (e.g., the Early Carboniferous and Late Cretaceous), when the Earth surface was covered mostly by oceans and seas, and the epochs of global regressions (e.g., the Late Permian-Triassic and the late Oligocene-early Miocene), when vast land areas existed.

Quite different approaches for the definition of sea-level and shoreline position were used in seismostratigraphic studies. Vail and others (Seismic Stratigraphy ..., 1977; Vail, 1987; Vail *et al.*, 1987) paid particular attention to the analysis of base (coastal) discordance of the layers (Fig. 1), which allows defining the sea level and tracing the paleoshorelines. They showed that this approach results in the definition of real sea-level changes. However, only a large amount of statistical data from various sedimentary basins can provide the opportunity of passing from the relative hypsographic

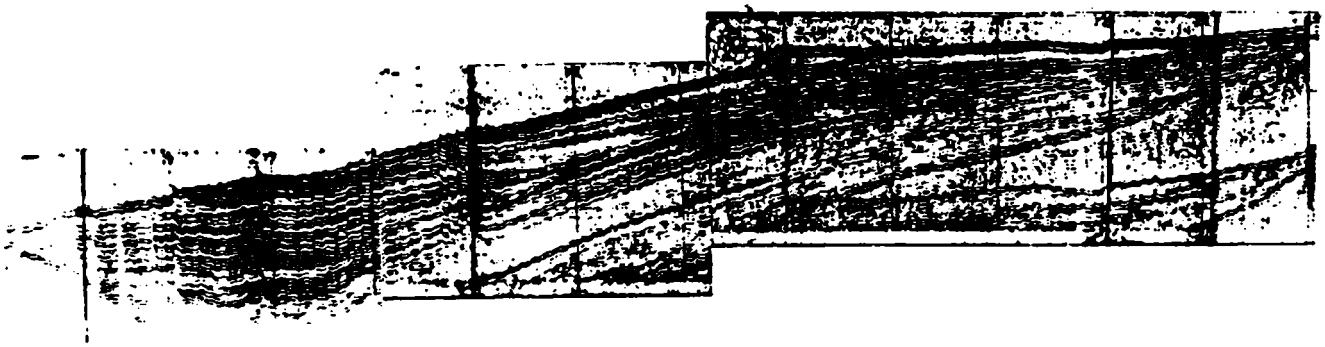


Fig. 1. *Fragment of the time section of the Primor'e shelf (the Japan Sea), demonstrating base (coastal) discordance of the Quaternary sediments.*

changes in the basin bottom and water surface to the fluctuations of the World Ocean level proper.

Base discordance allows tracing relative sea-level changes expressed in littoral facies. However, when analyzing continental or relatively deep-water facies, corrections should be introduced into estimations. In the first case, values of sea-level location should be subtracted from paleoshoreline hypsometry values, whereas in the second, they should be added to it. Vail and others stressed the necessity of introducing corrections into the estimates, taking into account the differences in sediment thickness caused by an uneven subsidence. Corrections are also needed when considering changes in the initial thickness of sedimentary sequences because of the decrease in rock volume in the course of their diagenesis.

Analysis of seismostratigraphic data shows that every cycle of relative sea-level change usually consists of its gradual rise (3 - 5 Ma), then a relatively short-term stabilization (some hundreds of ka), followed by a fast (momentary, in the geological sense) drop. Sea-level rise can be traced by a landward migration of the base (coastal) discordance; stabilization is marked by a toplap. Finally, sea-level fall results in a seaward migration of base (coastal) discordance. Smooth and long-term relative rises and sharp short-term falls of the sea level are, evidently, controlled by tectonic movements that change the ocean capacity. Descending tectonic movements are often of impulsive short-term character (Shlezinger, 1990). Unfortunately, base (coastal) discordances are rarely preserved in sequences (Gladenkov *et al.*, 1985; Shlezinger, 1990). The periphery of sedimentary basins, where base (coastal) discordances were formed, was subjected, as a rule, to ascending tectonic movements and intense erosion.

A eustatic cycle of sea-level fluctuation forms a sedimentary complex that is bounded with inconformities or has conformable boundaries. Sediment layering of varying type and range can be observed within the sedimentary complex: a low sea level, a trangression, and a high sea level. Low sea level is accompanied by intense lateral sedimentation with a clinoform formation, a process known as avalanche sedimentation.

There are other universal seismostratigraphic approaches that allow defining the sea-level position and tracing its relative changes. They include the analysis of clinoforms, whose formation is controlled by the sea-level position. When it is low and the shelf becomes a transit area, clinocyclites (regressive clinoforms) are formed, whereas the high level is marked by a generation of wedge covers (trangressive clinoforms) (Mikhailov and Shlezinger, 1989; Gogonenkov *et al.*, 1988; Igoshkin and Shlezinger, 1990a, b). Transitional zones between shelf and continental facies are used for qualitative estimations of relative sea-level changes (Kunin *et al.*, 1989). These zones allow defining paleoshorelines and tracing their migration and, as a result, objectively estimating relative sea-level changes. A facies analysis of the shelf sequences yields information on the sea-depth changes, while underwater deltas, bars, and erosion scours on land allow defining objectively the position of the sea paleosurface. If, for instance, relatively deep-sea deposits appear in the shelf sequence, they can be interpreted as evidence of a sea-level rise. Therefore, well-known eustatic curves have been compiled, mainly using data on facies analysis of sedimentary cycles (cyclothems) with their deep- and shallow-water parts defined. On a regional scale, such an approach is very fruitful. For instance, the Pliocene sequences of the northeast Kamchatka Peninsula record very short-term sea-level changes (at least three transgressions in a period of 3 Ma), yielding a key to understanding the history of the Bering Strait evolution (Fig. 2) (Gladenkov *et al.*, 1991). Comparison of cyclothems on a regional scale often reveals, however, significant discrepancies. For example, in some cases, deep-water facies in certain sequences correlate in time with regional regressions, and vice versa.

EUSTATIC SEA-LEVEL CHANGES

Two principal factors are responsible for eustatic or global sea-level changes (Naidin, 1971, 1976, 1985; Naidin *et al.*, 1981). The first of them is thought to be related to changes in the ocean-water volume and is controlled mainly by ice volume and the supply of juvenile water. Both glaciations and warmings can noticeably change the water volume in the ocean, which was most strikingly pronounced throughout the Oligocene-Quaternary (ice "takes" water, whereas warming "gives it back"). The second factor relates to the changes in the World Ocean capacity. The growth of mid-oceanic ridges and the evolution of deep geosyn-

clinal basins (their closing, due to folding, results in a sea-level rise, whereas the appearance of new oceanic basins leads to its fall) are thought to be the main causes of changes in ocean capacity. Other processes, such as climate changes, escape of water into space, changes in water properties, and some others, are of secondary importance (Kaplin, 1986). Eustatic sea-level changes create distinct marking boundaries of a subglobal scale. Vail and coauthors (Vail, 1987; Vail *et al.*, 1987) use these boundaries for delimiting the sedimentary complexes, considering the latter, as the main stratigraphical units of a sedimentary cover sequence.

TECTONIC MOVEMENTS RESULTING IN RELATIVE SEA-LEVEL CHANGES

Regional tectonic movements affecting the open sedimentary basins or their parts, i.e., basins connected with the World Ocean system, condition only relative sea-level changes that are manifested on a geographically limited scale. They change the basin depths and its shoreline, while the global sea level remains stable. These tectonic movements are usually characterized by insignificant amplitude (a few meters to a few tens of meters) and a weak differentiation, affecting relatively wide areas (many hundreds of kilometers). They do not form structural patterns of the basin, being superimposed on it, but they predetermine the position of the shoreline. Tectonic movements resulting in relative sea-level changes are often considered to be related to local isostasy. Thus, tectonic movements of the West Siberian platform in the Neocomian that resulted in relative sea-level fluctuations were caused by the isostatic loading of clinoforms of the same age (Igoshkin and Shlezinger, 1989a, b). High-amplitude vertical orogenic movements forming mountain belts predetermine relative sea-level falls and regressions. For instance, the draining of the vast areas of the Paleozoic Ural-Mongol fold belt and the retreat of the shoreline toward adjacent cratons were caused by orogenic movements during the Oligocene-Quaternary. Rifting with subsequent sea-floor spreading and short-term subsidences are the processes responsible for the initiation of new sea basins and pertinent shorelines. Thus, the Cenozoic shorelines of marginal seas along the Asian continental margin were formed as a result of an influence of these processes (Atlas ... , 1992).

The sea-level changes related to the changes in hypsometry of the basin bottom can coincide with eustatic fluctuations, intensifying their effect. In such cases, shoreline migrations of maximum amplitude are observed. In contrast, their opposite directions reduce the summary effect. When the former component surpasses the latter, anomalous regional deviations can arise.

STRUCTURE-FORMING TECTONIC MOVEMENTS

Tectonic movements forming the structural pattern of sedimentary basins play a more significant role in relative sea-level changes. Descending tectonic movements form a sedimentary cover and cause its thickness differentiation (Fig. 3). Ascending and lateral tectonic

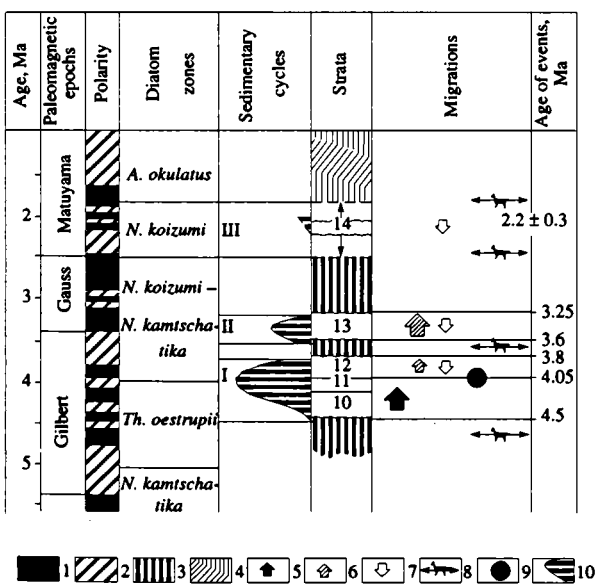


Fig. 2. *Chronology of Miocene geological events in northeastern Kamchatka.*

1 - 2 – paleomagnetic polarities: 1 – normal, 2 – reverse; 3 – hiatuses; 4 – angular unconformities; 5 - 7 – migrations of mollusks: 5 – low-boreal into the Bering Sea, 6 – of the Pacific origin into the Arctic Ocean and North Atlantic, 7 – of the North Atlantic origin into the Pacific; 8 – migrations of land mammals; 9 – initial opening of the Bering Strait; 10 – marine transgressions.

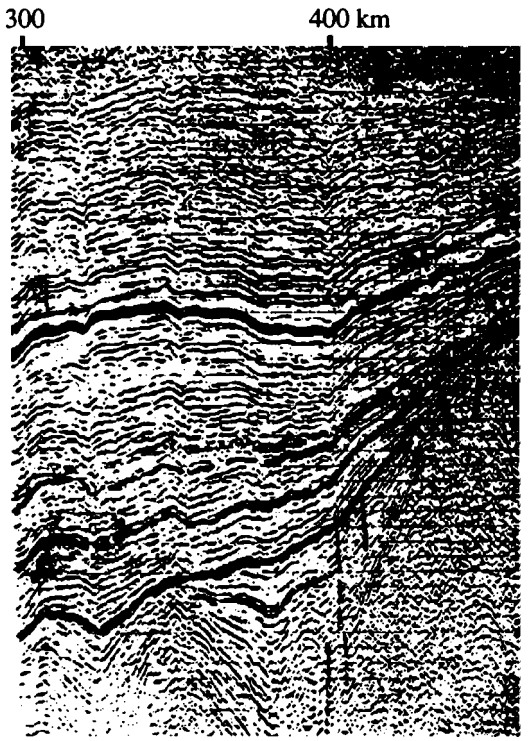
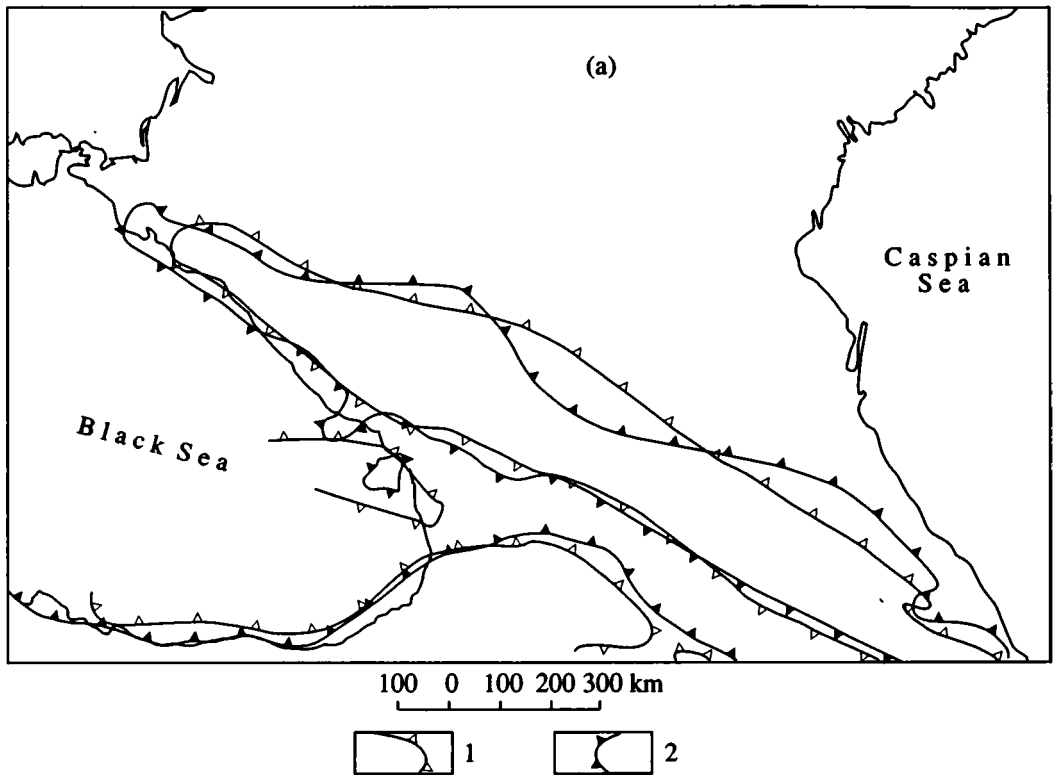
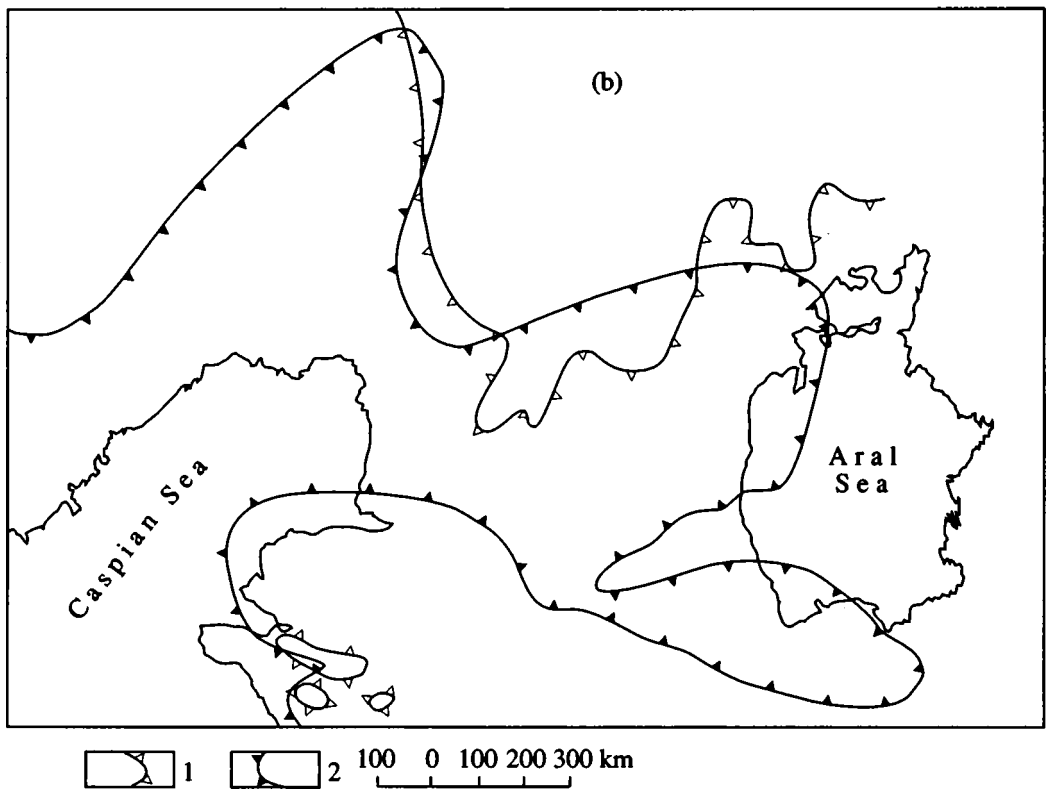


Fig. 3. *Fragment of the time section of the eastern slope of the West Siberia platform demonstrating synsedimentary tectonic subsidence.*


(a)
Caspian Sea
Black Sea
100 0 100 200 300 km
1
2


(b)
Aral Sea
Caspian Sea
1
2
100 0 100 200 300 km

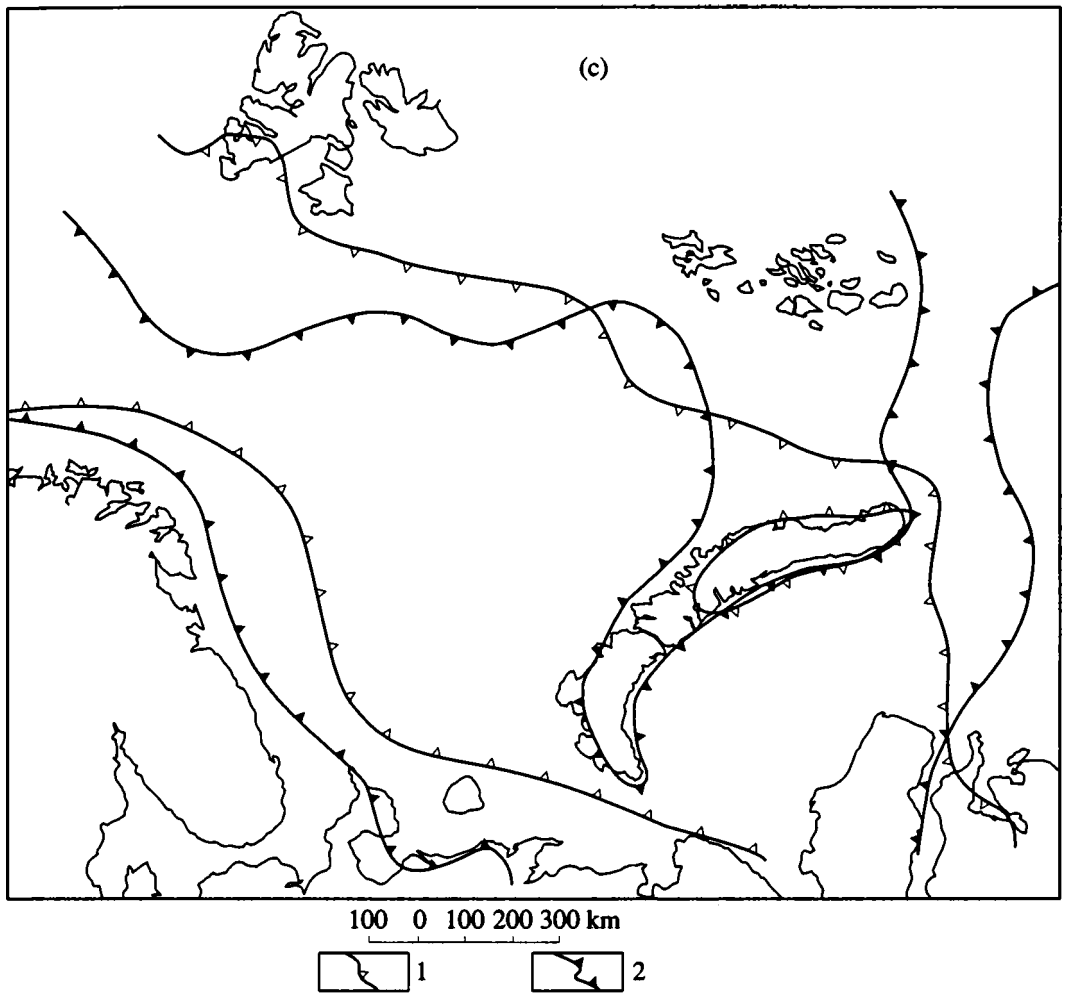


Fig. 4. *Examples of paleogeographic situations (a - c) (Atlas ... , 1992).*
a – Great Caucasus; b – Northern Aral and Caspian areas. Paleoshorelines: 1 – in the Oligocene, 2 – in the middle Miocene; c – the Barents and Kara seas. Paleoshorelines: 1 – in the Early Cretaceous, 2 – in the Late Cretaceous. Notches are directed seaward.

movements result in erosion of the sedimentary cover. Structure-forming tectonic movements do not significantly change the sea depth or position of the shorelines. Deposits of maximum thickness were formed in the areas of the greatest synsedimentary subsidence, whereas a decrease of the subsidence resulted in the sediment thickness tapering to a point. Thus, synsedimentary descending movements are mainly responsible for the formation of the sedimentary basin structural pattern. Their zero values define sedimentation limits rather than a shoreline. Wide shelf-floor areas can be situated behind this line, where no sedimentation took place. Therefore, the synsedimentary wedging of layers cannot always be used as a criterion for paleogeographic reconstructions, and, unfortunately, it is sometimes overused (Gavrilov, 1989). Exceptions are steep areas of land conjugated with the shelf. In these areas, despite the subsidence, no sedimentation took place, and terrigenous clastic material was transported to the shelf. In this case, the synsedimentary wedging coincided with the base (coastal) discordance (Zabolotnikov, 1989; Shlezinger, 1989).

Ascending vertical and lateral compression movements in the shelf area are accompanied by underwater erosion of the uplifting bottom. Because of wave erosion, the shelf bottom remains at the same level, while the rocks composing it are destroyed, and erosional surfaces, after their subsequent burial by sediments, form angular (structural) unconformities.

Tectonic elements created by structure-forming movements often do not coincide with the shoreline and may sometimes cross it. In this case, the thicker sedimentary cover can be formed on the land, with the maximum erosion confined to the shelf. Only the very intense ascending vertical tectonic movements (orogenesis), estimated in hundreds of meters to several kilometers in the time interval but not exceeding several hundred ka, result in the failure of underwater erosion to destroy the uplifting bottom, which leads, as was shown above, to a relative sea-level fall and a change in shoreline position. Synsedimentary descending movements predetermine the sedimentary cover thickness in epicontinental basins. At the same time,

changes in water depth affect mainly the facies composition of deposits.

The synsedimentary wedging of layers in response to descending tectonic movements resembles in form a base (coastal) discordance, which is formed by sea-level changes. The divergence of layers with insignificant or completely absent facies changes is the most distinctive feature of the former. In contrast, a base (coastal) discordance is characterized by parallel layering with a distinct facies transition from continental through coastal to outer-shelf deposits.

SEDIMENT SUPPLY AND SEQUENCE CYCLES

The delivery of terrigenous clastic material into the basin results in a relative sea-level fall and a seaward migration of the shoreline. This is particularly significant in the shelf areas adjacent to river deltas. For instance, in the deltas of such large rivers as the Mississippi, Ganges, Lena, and others, the sediment supply is responsible for seaward migrations of the shoreline of tens of kilometers, giving the impression of a local regression (Atlas ... , 1992).

Vail's studies (Seismic Stratigraphy ... , 1977) showed that transgressive and regressive cycles in sequences do not necessarily coincide with relative sea-level rises and falls, respectively, or with shoreline migrations (transgression and regression). On a local scale, transgressive and regressive successions in the sequences and lithofacies migrations depend, first, on the volume of the terrigenous clastic material supply into the basin, whereas the global sea-level changes relate to eustasy and non-structural tectonic movements. During the periods of a relative sea-level rise (transgression) and a high input of terrigenous clastic material, the coastal facies are displaced seaward, which is reflected in the regressive succession of sediments in the sequence. On the other hand, when the sea level falls (regression) and the supply of terrigenous material diminishes, the marine facies can migrate landward and the sequence will be characterized by a transgressive succession. A comparison of sedimentary cycles, which were established in many sequences in Russia using a facies analysis, with a eustatic curve compiled according to base (coastal) discordances, confirms Vail's conceptions. For instance, according to I.N. Karogodin (1980), the boundaries of transgressive mesocyclites correlate with the middle Callovian and middle Albian in western Siberia and with the late Paleocene in the Maili Sai section in Kirgizia, whereas on the eustatic curve, these events are missing.

PALEOGEOGRAPHIC RECONSTRUCTIONS

Some earlier works postulated that there are significant chronological discrepancies in the distribution of transgressions and regressions on different continents. For instance, A.L. Yanshin (1973), who analyzed their spatial distribution in detail, came to the conclusion that they were controlled by vertical movements of major blocks of the Earth's crust, rather than by global eustatic sea-level changes. Many of Yanshin's ideas, however, were later subjected to criticism (Kazarinov, 1976). Indeed, in many regions, relative sea-level fluctuations, caused by tectonic movements and terrigenous clastic material supply, exceeded the effect of eustatic fluctuations, which resulted in local transgressions and regressions. The analysis of paleogeographic maps of the Eurasian Mesozoic and Cenozoic shelves (Atlas ... , 1992) provides much evidence confirming these observations. For example, the orogenic growth of the Caucasus in the middle-late Miocene was accompanied by a seaward shoreline migration, though a global sea-level rise took place at that time (Fig. 4a). A similar situation can be observed in the northern Aral area and in the eastern Caspian basin: the area of middle-upper Miocene marine deposit distribution was, in some places, significantly reduced as compared with that in the late Oligocene (Fig. 4b). The water area of the Kara Sea basin was also significantly reduced from the Early to Late Cretaceous, despite the fact that this period coincided with a global sea-level rise (Fig. 4c). Further examples of such discrepancies between shoreline migrations and eustatic curves could be listed. They clearly demonstrate that transgressions and regressions have been controlled in many regions by tectonic movements rather than by eustatic sea-level rises and falls. Tectonic movements are of particular importance in geosynclinal basins transforming into mountain foldbelts. Shorelines of zones of the mountain foldbelts, in conjunction with cratons, have been controlled solely by ascending vertical and compressional lateral tectonic movements. The role of the terrigenous material supply has been less significant.

CONCLUSION

The data discussed shows a lack of a direct correlation of transgressions and regressions with the global sea-level position or, at least, it shows their complex interrelationship. Regional tectonic movements that cause relative sea-level changes and a terrigenous clastic material supply can significantly complicate the effect of global eustasy. When these processes are of a large scale and act in an opposite direction with respect to eustasy, the transgressions and regressions in certain areas can disagree with the global cycles of sea-level changes. At the same time, the eustasy left a distinct impact on the geological record, first of all, on the structure of sedimentary complexes (cyclothems). Reconstruction of eustatic fluctuations offers a key to understanding their nature. It would be proper to consider them as very important in the formation of sedimentary basins.

REFERENCES

Atlas Litologo-Paleogeograficheskikh Kart SSSR (Atlas of Lithologic-Paleogeographic Maps of the USSR), Vinogradov, A.P., Ed., Aerogeologiya, Moscow, 1967, vol. 4; 1968, vols. 1, 3; 1969, vol. 2.

Atlas of the Paleogeography of China, Cartographic Publishing House, Beijing, China, 1985.

Atlas Paleogeograficheskikh Kart. Shel'fy Evrazii v Mezozoe i Kainozoe. Tom 2. Karty (Atlas of Paleogeographic Maps. Eurasian Shelves in the Mesozoic and Cenozoic. vol. 2. Maps), Robertson Group and Geological Institute of the Russian Academy of Sciences, 1992.

Gavrilov, Yu.O., Sedimentation in the Great Caucasus Early-Middle Jurassic Basin (in Connection with Transgressions and Regressions), *Litologiya i Poleznye Iskopaemye*, 1989, no. 2, pp. 3 - 15.

Gladenkov, Yu.B., Barinov, K.B., and Basilyan, A.E., Detailed Stratigraphy of the Eastern Kamchatka Pliocene, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1991, no. 9, pp. 79 - 86.

Gladenkov, Yu.B., Kunin, N.Ya., and Shlezinger, A.E., Seismic Stratigraphy Abroad, *Byull. MOIP, Otd. Geol.*, 1985, vol. 60, no. 5, pp. 100 - 105.

Igoshkin, V.P. and Shlezinger, A.E., Genetic Types of the Neocomian Clinoforms in West Siberia, *Geol. Geofiz.*, 1990, no. 8, pp. 16 - 20.

Igoshkin, V.P. and Shlezinger, A.E., Neocomian Clinoforms in West Siberia and Causes of Their Origin, *Dokl. Akad. Nauk SSSR*, 1990, vol. 312, no. 4, pp. 1191 - 1195.

Kaplin, P.A., Types of Sea-Level Changes, *Geomorfologiya*, 1986, no. 9, pp. 16 - 22.

Karagodin, Yu.N., *Sedimentatsionnaya Tsiklichnost'* (Sedimentary Cyclicity), Nedra, Moscow, 1980, p. 242.

Kazarinov, V.P., On the Way of Deciphering Paleogeographic Regularities in the Earth's Evolution, *Byull. MOIP, Otd. Geol.*, 1976, vol. 51, no. 2, pp. 5 - 32.

Kunin, N.Ya, Kosova, S.S., and Blokhina, G.Yu., *Prognozirovanie Neantiklinal'nykh Lovushek Nefti i Gaza na Osnove Seismostratigraficheskogo Analiza (na Primere Vostochnogo Predkavkaz'ya)* (Prognostication of Non-Anticlinal Traps of Oil and Gas Using Seismostratigraphic Analysis (With Reference to the Eastern Part of the Caucasian Northern Foothills)), VNIIOEN, Moscow, 1989, p. 42.

Metody Sostavleniya Litologo-Fatsial'nykh i Paleogeograficheskikh Kart (Methods of Compiling Lithofacies and Paleogeographic Maps), SO Akad. Nauk, Novosibirsk, 1963, p. 176.

Mikhailov, Yu.A. and Shlezinger, A.E., Relative Sea-Level Changes, *Sov. Geol.*, 1989, no. 11, pp. 68 - 73.

Naidin, D.P., Epeirogenesis and Eustasy, *Vestn. Mosk. Gos. Univ., Geol.*, 1976, no. 2, pp. 3 - 16.

Naidin, D.P., On the Sea-Level Changes in the Mesozoic and Cenozoic, *Byull. MOIP, Otd. Geol.*, 1971, vol. 41, no. 3. pp. 10 - 18.

Naidin, D.P., World Ocean Level at the End of the Mesozoic, *Vestn. Mosk. Gos. Univ., Geol.*, 1985, no. 3, pp. 10 - 18.

Naidin, D.P., Sazonova, I.G., Poyarkova, Z.N., *et al.*, Cretaceous Transgressions and Regressions in the East European Craton, Crimea and Central Asia Regions, *Byull. MOIP, Otd. Geol.*, 1980, vol. 55, no. 5, pp. 27 - 42.

Schuchert, Ch., *Paleogeograficheskii Atlas Severnoi Ameriki* (Paleogeographic Atlas of Northern America), Inostrannaya Literatura, Moscow, 1957.

Seismic Stratigraphy – Application to Hydrocarbon Exploration, Mem. 26 AAGP, Tulsa, Oklahoma, USA, 1977.

Shlezinger, A.E., Explanations and Critical Notes to the Book Seismic Stratigraphy, *Seismostratigraficheskie Issledovaniya v SSSR*, Nauka, Moscow, 1990, pp. 30 - 39.

Shlezinger, A.E., Seismostratigraphy in Defining Relative Sea-Level Fluctuations, *Byull. MOIP, Otd. Geol.*, 1989, vol. 64, no. 1, p. 123.

Shlezinger, A.E., Tectonic Structures of the World Ocean Floor, *Byull. MOIP, Otd. Geol.*, 1990, vol. 65, no. 6, pp. 29 - 34.

Vail, P.R., Seismic Stratigraphy Interpretation Procedure. Atlas of Seismic Stratigraphy, *AAPG in Geology*, 1987, vol. 27, no. 1, pp. 1 - 10.

Vail, P.R., Colin, J.P., Chene, du R.J., *et al.*, La stratigraphie sequetielle et son application aux correlations chronostratigraphiques dans le Jurassique du Bassin de Paris, *Byull. Soc. Geol. France*, Ser. 8, 1987, vol. 111, no. 7, pp. 1301 - 1321.

Yanshin, A.L., On the So-Called World Trangressions and Regressions, *Byull. MOIP, Otd. Geol.*, 1973, vol. 48, no. 8, pp. 9 - 44.

Zabolotnikov, A.A., On the Possibilities of Defining Relative Sea-Level Fluctuations under Conditions of the Synsedimentary Subsidence, *Byull. MOIP, Otd. Geol.*, 1989, vol. 64, no. 1, p. 124.

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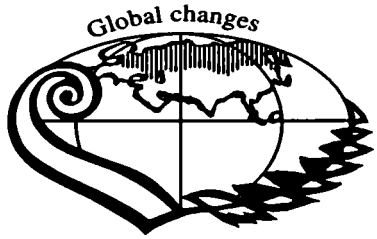

Evolution of the biosphere

Evolution of the Composition of Carbonate Rocks in the Late Precambrian

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Abstract – A generalization of the extensive body of data on the petrochemistry of carbonate rocks from the Riphean and Vendian of the Urals and Siberia leads to a conclusion of the geochemical specificity of Upper Riphean carbonate rocks that differ from the respective rocks of the under- and overlying strata in a lower dolomites/limestones ratio and a higher Sr content. The model below illustrates the evolution of carbonate accumulation, showing that the cause of progressively waning dolomite formation in the geologic history was an irreversible process of growing Ca concentrations in the Earth's sediments due to sea water interaction with oceanic basalts in hydrothermal systems. Global cyclic variations in the dolomites/limestones ratio may be a result of the respective changes in the rate of CO_2 input in the atmosphere.

Among the empirical data that served as a basis for the elaboration and development of the concepts of the evolution of sedimentation, the most crucial was a discovery suggesting that carbonate rocks, the most widespread products of authigenic mineral formation in the sedimentary cover, did change in composition with geologic history. The works by Daly (1909), Vinogradov and Ronov (1956), Veizer (1978), and others showed that Phanerozoic and Proterozoic deposits exhibited a tendency toward an increasing role of dolomites among the other carbonate rocks, along with an increasing age. An example of a thorough investigation of time-related changes of carbonate compositions was a paper by Vinogradov and Ronov (1956), which provided mean compositions of these rocks for the series or systems of the Russian platform Phanerozoic, thus suggesting not only a reliable trend, but also important cyclic variations of the Mg/Ca ratio in carbonate mean compositions throughout the Paleozoic. The paper, followed by other papers on this problem, discussed data on Precambrian carbonate rocks only in the most general form, or else had only a narrow, regional database. The paper attempts a more detailed analysis of the carbonate accumulation evolution over the billion years that preceded the Phanerozoic.

COMPOSITION OF THE UPPER PRECAMBRIAN CARBONATES

The analytical data used in this paper includes 1169 analyses of petrogenic elements in bulk compositions of samples of Riphean and Vendian carbonate rocks from the Urals (mostly the Bashkirian anticlinorium) and the Siberian platform and its fringes (Table 1). About half of those were obtained by us ourselves; the rest were made available by our colleagues from various geologic establishments in this country (E.Z. Gareev, B.R. Shpunt, V.K. Golovenok, V.K. Nemerov, E.A. Zubareva, and Ya.E. Yudovich) or were borrowed from the literature (Petrov, 1982).

Ca and Mg. Each of four samples of the Upper Precambrian rock analyses grouped according to their ages (Fig. 1) shows a bimodal distribution of MgO, with maxima between 0 - 2% and 19 - 22%. Such a pattern of MgO distribution has been noted earlier, particularly in Phanerozoic carbonates (Ronov, 1956; Carbonate ... , 1967b; Veizer, 1978). Thus, nature has a higher abundance of those carbonate rocks that have a fairly uniform mineral composition of the carbonate phase: either calcite or dolomite, while mixed rocks are relatively rare. In some formations of the studied carbonate sequences, carbonate strata may show either a predominantly limestone or dolomite composition, or they may be composed of alternating limestones and dolomites; in the latter case, mixed carbonates would also be restricted in distribution (Fig. 2).

Studies of modern carbonate sediments prove that major minerals that composed carbonate rocks, the low-magnesian calcite and, in particular, dolomite, are largely products of an early diagenetic alteration of meta-stable primary minerals typical of modern sediments: aragonite, high-magnesian calcite, and proto-

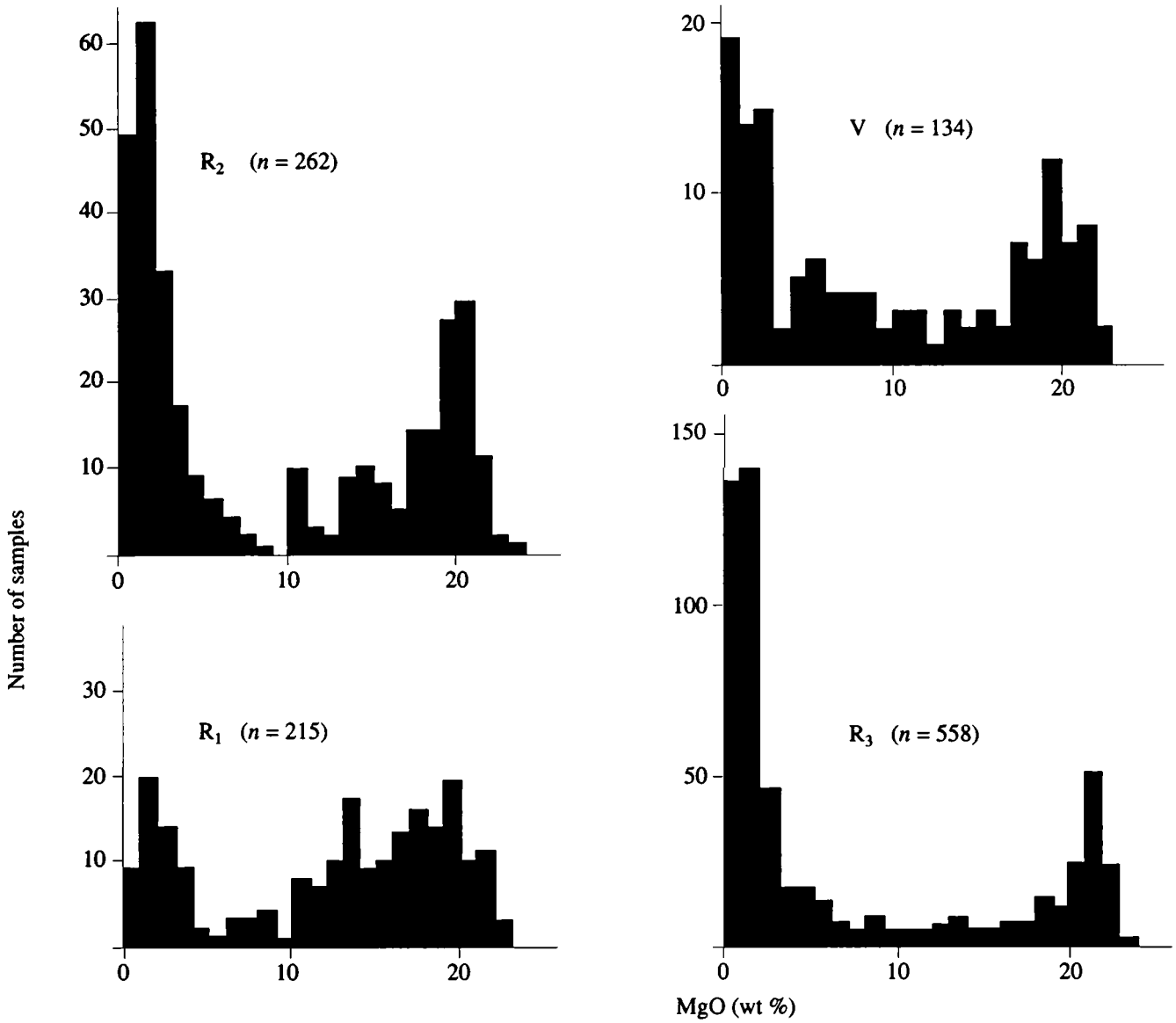

Fig. 1. *Histograms of MgO distribution in the Riphean and Vendian carbonate rocks in the regions studied.*

dolomite (for a review of the problem, see (Carbonate ... , 1967a, b; Carbonates ... , 1983; Kholodov, 1988)). On the other hand, there is reason to doubt that this diagenetic model of the formation of carbonate rock compositions can be applicable throughout the geologic history. Sedimentary environments different from modern ones and, in the first place, higher partial CO_2 pressures could have been responsible for the synse-dimentary accumulation of low-magnesian calcite and dolomites in older stages of the Earth's history (Pigott and Mackenzie, 1979; Mackenzie and Pigott, 1981). The epigenetic processes of limestone dolomitization at the katagenetic stage could also have been important for dolomite formation (Kholodov, 1988). The question of the role that these processes played in the composition of carbonate rocks is the subject of discussion.

The observed distribution of MgO content in Precambrian carbonate rocks (Fig. 1) can best be explained from the positions of the predominantly early diagenetic or synsedimentary origin of dolomite rocks. MgO contents that fall within the minima in the middle of the histograms correspond to the composition of high-magnesian calcite, a meta-stable mineral replaced under diagenesis by low-magnesian calcite or dolomite. The system then should be sufficiently open both for Ca and Mg. This condition conforms more to the early diagenesis rather than to later stages in the carbonate rock history. In terms of the synsedimentary model, the bimodal distribution of magnesium in carbonate rocks can be attributed to the restricted transition interval that separates fields of physico-chemical properties of the environment that control carbonate precipitation, either in the form of low-magnesian calcite, or as dolomite. If the origin of dolomites is approached from the point of

Table 1. Mean chemical composition of carbonate rocks from the studied late Precambrian formations of the Urals and Siberia

General scale	Formation	Number of analyses	SiO_2	TiO_2	Al_2O_3	FeO + Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	LOI	Total	Ba	Sr	Mg/Ca
Urals Fold Belt																	
Bashkirian meganticlinorium																	
Lower Riphean	Satka	69	15.44 11.04	0.09 0.08	2.21 1.77	1.46 1.07	0.03 0.03	15.93 4.65	26.02 5.41	0.14 0.20	0.92 0.88	0.03 0.03	37.38 6.59	99.65	57 24	38 48	0.85
	Bakal	21	8.45 7.25	0.08 0.06	2.07 1.74	4.52 4.08	0.16 0.13	18.14 6.80	24.59 5.80	0.24 0.10	0.56 0.35	0.02 0.02	41.35 5.29	100.18	132 172	212 159	1.03
	Sura	65	14.62 11.79	0.10 0.11	2.12 2.18	1.61 1.46	0.08 0.07	7.45 7.75	36.31 11.57	0.34 0.54	0.79 0.68	0.03 0.02	36.62 7.01	100.07	49 22	200 67	0.29
Middle Riphean	Avzyan	99	11.20 8.49	0.07 0.09	1.94 1.48	2.00 1.92	0.10 0.14	13.14 7.39	31.75 8.40	0.10 0.13	0.58 0.53	0.03 0.04	39.05 5.00	99.96	177 612	202 170	0.58
Upper Riphean	Zil'merdak	10	28.74 8.56	0.30 0.12	5.86 2.03	4.31 0.79	0.16 0.09	12.13 3.26	18.71 4.29	0.71 0.23	1.78 0.59	0.08 0.03	27.09 5.18	99.87	130 107	114 97	0.90
	Katav	68	11.64 6.35	0.12 0.11	2.46 1.48	1.78 1.75	0.07 0.04	1.98 2.04	44.43 6.31	0.31 0.23	0.53 0.34	0.05 0.03	36.48 3.93	99.85	155 482	498 441	0.06
	Inzer	48	4.11 4.03	0.04 0.04	0.78 0.74	0.55 0.38	0.06 0.04	1.30 1.40	51.07 3.62	0.24 0.22	0.22 0.16	0.03 0.03	41.07 2.43	99.47	157 835	1006 976	0.04
	Min'yar	94	3.78 4.74	0.03 0.03	0.58 0.53	0.51 0.48	0.05 0.07	18.78 6.84	32.06 6.54	0.17 0.12	0.19 0.23	0.02 0.02	43.49 3.15	99.66	39 34	229 141	0.81
	Uk	59	9.86 10.69	0.12 0.15	1.90 2.62	1.68 1.24	0.07 0.08	3.35 4.96	44.17 9.09	0.19 0.19	0.58 0.76	0.03 0.03	37.90 6.31	99.85	67 82	234 166	0.11
Lyapino anticlinorium																	
Upper Riphean	Moroin	5	13.22 8.71	0.12 0.05	2.07 0.80	1.03 0.43	0.18 0.14	1.25 1.90	45.38 5.46	0.30 0.39	0.35 0.24	0.11 0.12	36.19 4.45	100.20	–	–	0.04

Table 1. Contd.

General scale	Formation	Number of analyses	SiO_2	TiO_2	Al_2O_3	FeO + Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	LOI	Total	Ba	Sr	Mg/Ca
Pai-Khoi anticlinorium																	
Vendian	Yasarusalinsk	9	14.07 8.71	0.29 0.37	3.42 2.25	2.35 1.36	0.37 0.51	7.56 8.33	35.57 13.34	0.41 0.32	0.41 0.42	0.13 0.08	34.59 5.68	99. 17	–	–	0.30
Siberian Platform																	
Ura anticlinorium																	
Upper Riphean	Barakun	8	26.47 16.73	0.26 0.21	5.15 3.44	2.15 1.14	0.03 0.02	3.10 3.09	31.17 13.50	1.16 0.72	0.64 0.47	0.07 0.11	28.79 9.77	98.99	556	1492	0.14
	Chencha	37	9.26 8.11	0.08 0.09	1.65 1.40	0.98 0.99	0.04 0.05	3.55 4.53	44.78 9.10	0.20 0.18	0.46 0.51	0.04 0.04	38.46 4.94	99.50	–	786	0.11
Vendian	Tinnaya	38	6.00 6.20	0.07 0.09	1.22 1.51	0.96 1.44	0.24 0.54	6.13 6.92	43.61 10.94	0.39 0.21	0.25 0.35	0.05 0.05	40.91 3.86	99.83	–	270	0.20
Uchur-Maya plate																	
Middle Riphean	Malgina	23	8.42 3.57	0.06 0.04	2.56 0.73	1.23 0.82	0.06 0.06	2.98 4.64	45.16 6.36	0.12 0.19	0.29 0.52	0.02 0.01	38.64 2.26	99.54	131 180	122 33	0.09
	Tsipanda	26	2.76 1.70	0.02 0.01	0.66 0.39	0.72 0.30	0.01 0.01	19.53 1.23	30.69 1.46	0.01 0.01	0.07 0.07	0.04 0.03	45.18 1.12	99.69	39 11	71 61	0.88
Upper Riphean	Ignikan	19	2.12 1.86	0.02 0.03	1.12 0.87	0.90 1.03	0.03 0.03	6.56 6.24	45.44 7.96	0.01 0.01	0.05 0.01	0.02 0.01	43.26 1.87	99.53	55 59	142 130	0.20
	Neryuen	14	4.48 2.69	0.05 0.06	1.70 1.03	2.20 1.66	0.06 0.05	10.84 7.13	38.11 8.33	0.09 0.13	0.11 0.14	0.02 0.01	41.92 2.83	99.58	59 43	67 30	0.40
Vendian	Aim + Ust'-Yudoma	12	13.27 11.65	0.01 0.01	0.78 0.44	0.37 0.22	0.01 0.01	16.97 2.54	28.20 4.42	0.04 0.03	0.08 0.10	0.03 0.02	40.09 5.24	99.85	39 12	42 9	0.84

Table 1. Contd.

General scale	Formation	Number of analyses	SiO_2	TiO_2	Al_2O_3	FeO + Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	LOI	Total	Ba	Sr	Mg/Ca
Marginal structures of the Siberian Platform and the folded fringes																	
Yudoma-Maya depression																	
Lower Riphean	Trekhgorka	16	22.28 12.43	0.08 0.08	3.03 2.12	2.01 1.79	0.05 0.06	14.41 3.34	21.75 5.17	0.29 0.37	1.26 0.96	0.04 0.03	34.29 7.03	99.49	199 167	60 36	0.92
	Dim	20	21.78 15.36	0.09 0.10	2.90 2.32	2.63 1.67	0.06 0.05	14.49 2.97	22.32 6.15	0.09 0.17	1.54 1.55	0.05 0.04	33.62 8.75	99.57	342 374	59 15	0.90
Middle Riphean	Svetlyi	7	16.18 7.62	0.06 0.08	1.54 1.07	1.87 1.71	0.08 0.08	15.87 1.87	25.21 3.08	0.10 0.07	0.18 0.25	0.03 0.03	38.39 4.38	99.51	43 8	99 29	0.88
	Malgina	13	8.52 10.43	0.04 0.03	0.93 0.54	0.73 0.46	0.02 0.02	7.77 7.72	40.64 10.00	0.07 0.07	0.10 0.11	0.05 0.04	40.70 4.61	99.57	109 183	229 246	0.27
	Tsipanda	22	8.00 8.84	0.02 0.02	0.95 0.33	0.44 0.35	0.01 0.01	14.09 7.56	34.12 10.02	0.02 0.02	0.08 0.10	0.03 0.03	41.81 3.78	99.57	40	104 134	0.57
Upper Riphean	Ignikan	43	5.45 6.73	0.08 0.14	2.56 2.17	1.43 1.92	0.05 0.06	4.92 5.70	44.28 11.64	0.05 0.05	0.29 0.51	0.02 0.01	40.42 4.21	99.55	58 74	311 225	0.15
	Neryuen	9	9.34 7.45	0.15 0.19	4.16 3.12	4.58 4.54	0.11 0.09	7.46 4.16	35.48 11.15	0.09 0.08	0.45 0.69	0.02 0.01	37.73 6.26	99.57	67 52	349 256	0.29
Vendian	Ust'-Yudoma	8	7.48 8.46	0.05 0.10	2.16 1.80	0.80 1.07	0.03 0.03	4.89 2.17	44.00 7.95	0.14 0.25	0.15 0.28	0.02 0.01	39.87 5.40	99.59	45 14	405 431	0.15
Sette-Daban horst-anticlinarium																	
Vendian	Yalan	9	11.47 10.22	0.06 0.05	2.43 1.21	0.88 0.59	0.04 0.04	9.64 5.29	38.62 9.24	0.25 0.22	0.28 0.17	0.07 0.02	36.47 8.15	100.21	228 209	401 356	0.35
	Sardana	15	4.11 3.42	0.04 0.02	1.56 0.33	0.42 0.30	0.04 0.02	12.73 8.88	38.97 10.54	0.16 0.16	0.25 0.22	0.24 0.16	42.62 2.94	101.14	–	752 1157	0.45

Table 1. Contd.

General scale	Formation	Number of analyses	SiO_2	TiO_2	Al_2O_3	FeO + Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	LOI	Total	Ba	Sr	Mg/Ca
Turukhansk uplift																	
Middle Riphean	Sukhaya Tunguska	22	8.06 6.54	0.07 0.03	0.61 0.51	1.42 1.38	0.03 0.04	10.09 8.67	38.24 10.70	0.09 0.05	0.29 0.31	0.03 0.01	40.55 4.00	99.48	–	–	0.37
Upper Riphean	Derevnya	9	5.60 2.87	0.09 0.03	1.24 0.75	3.46 2.00	0.11 0.10	17.78 3.68	29.09 3.51	0.07 0.04	0.37 0.21	0.02 0.01	41.92 2.50	99.75	–	–	0.85
	Nizhnyaya Tunguska	12	2.53 3.14	0.05 0.04	0.41 0.37	0.58 0.38	0.03 0.04	5.53 7.87	47.79 8.57	0.06 0.08	0.15 0.10	0.05 0.08	42.79 2.38	99.97	–	–	0.16
	Burovoi	8	5.49 7.75	0.03 0.02	0.10 0.10	1.17 0.43	0.02 0.01	20.70 1.88	28.70 2.52	0.05 0.01	0.07 0.04	0.04 0.04	43.73 3.68	100.10	–	–	1.00
	Durnoi Mys	10	5.61 4.08	0.10 0.06	1.13 1.00	2.06 0.67	0.02 0.01	19.59 1.76	28.28 1.50	0.05 0.03	0.34 0.32	0.03 0.01	42.66 2.51	99.87	–	–	0.96
Enisei Ridge																	
Lower Riphean	Korda	9	26.74 8.87	0.16 0.13	3.00 1.79	3.98 3.46	0.32 0.41	8.90 5.99	26.83 6.59	0.52 0.63	0.49 0.26	0.06 0.05	28.84 7.32	99.84	–	–	0.46
Igarka uplift																	
Upper Riphean	Chernaya Rechka	25	6.08 5.56	0.07 0.03	1.00 0.48	1.07 0.71	0.02 0.01	0.97 0.58	49.99 3.57	0.14 0.06	0.28 0.14	0.04 0.05	40.23 2.72	99.89	–	–	0.027
Baikal complex																	
Upper Riphean	Goloe Ust'ye	9	17.92 13.09	0.11 0.05	1.88 1.06	1.48 0.92	0.11 0.05	12.89 5.59	29.11 8.24	0.07 0.03	0.35 0.31	0.07 0.05	35.49 7.72	99.48	44 12	110 50	0.62
	Uluntui	11	9.57 8.74	0.06 0.06	1.38 1.24	0.87 1.28	0.05 0.05	7.87 4.48	39.82 10.57	0.11 0.09	0.22 0.25	0.07 0.03	40.00 4.78	100.02	42 30	220 139	0.27

Notes: All oxides in wt %, Ba and Sr in ppm; mean arithmetic in numerator, standard deviation in denominator. Listed are formations for which at least five complete chemical analyses of petrogenic elements in the carbonate rocks are available. Formations are assigned to the general scale's subdivisions according to (Semikhatov and Serebryakov, 1983; Stratotip rifeya..., 1983), and for Southern Siberia, (Khomentovskii, 1990).

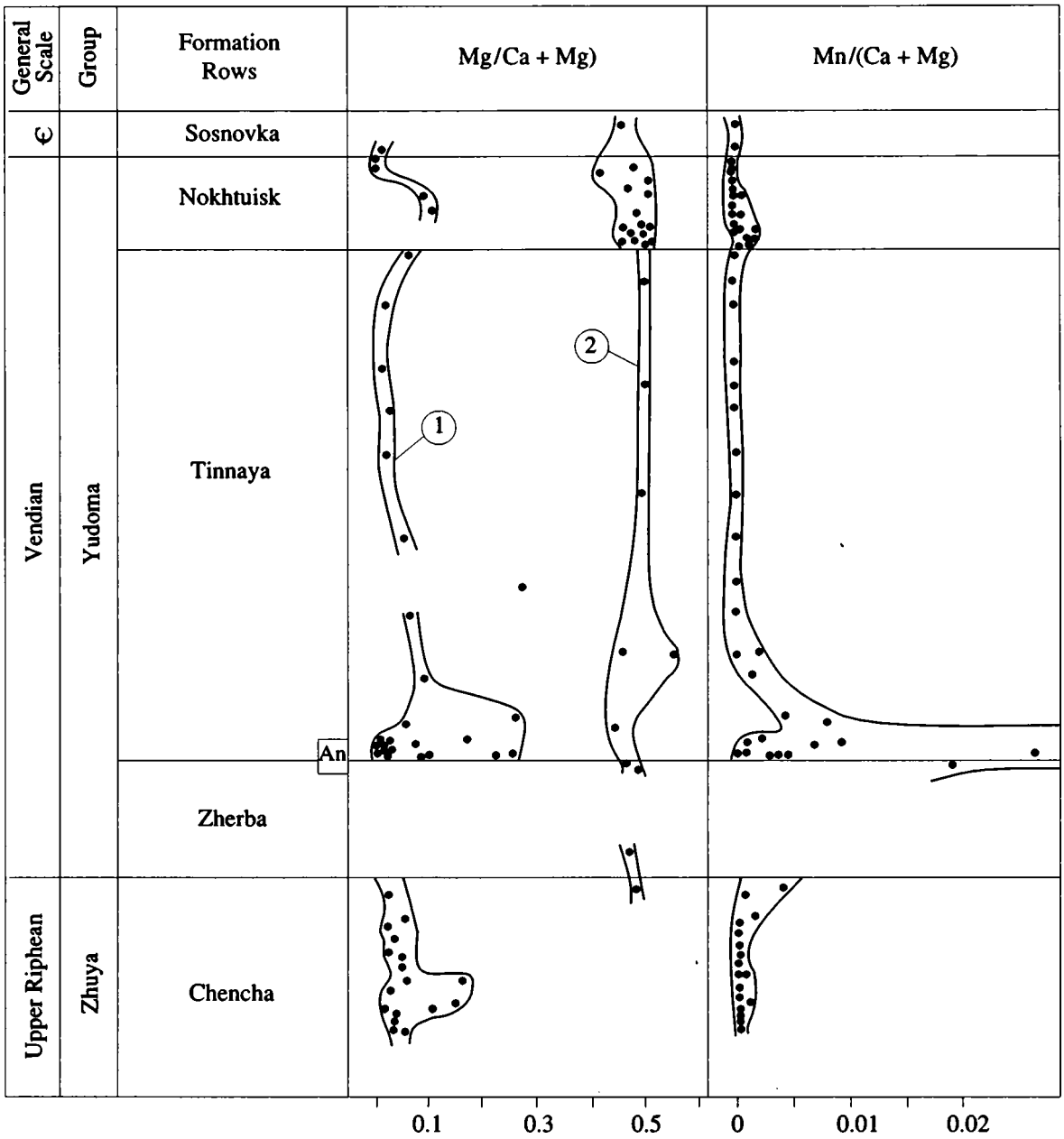

Fig. 2. *Variations of petrochemical modulus (atomic quantities) for carbonate rocks in the Upper Riphean, Vendian and Cambrian sections on the Lena River, near Nokhtuisk, Ura anticlinorium, southern Siberian platform.*
1 – limestones, 2 – dolomites.

view of the predominant epigenetic control at the katagenesis stage, it is difficult to explain why deep-seated waters altered calcareous rocks largely in rather pure dolomites, while the intermediate stage of partial dolomitization is fairly rare.

Comparisons of the histograms and data from Table 2 suggest the basic trend of changes in the dolomite/limestone ratios in carbonate rocks from the studied regions in the course of Late Precambrian history. From the Early to Late Riphean, the proportion of dolomites falls, then increases again in the Vendian. This general trend can also be observed in some major structural units (Fig. 3), with a relatively wide spread of mean compositions of carbonate rocks in the Middle Riphean.

The relatively limited distribution of dolomites in the Upper Riphean carbonates, in comparison with the older unit, is typical not only of the Urals or Siberia: it has also been recorded in the Upper Proterozoic sequences of the North China and Yangtze platforms. There, carbonate rocks in the Jingeryu and Teling formations in the former region, and Wangangxi in the latter, correlated with the Upper Riphean (Semikhatov, 1974), are represented almost entirely by limestones, whereas in the middle and lower portions of the

Table 2. Average compositions of carbonate rocks from late Precambrian units in the studied regions

Stratigraphic units	Number of analyses	SiO_2	TiO_2	Al_2O_3	FeO*	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Ba	Sr	LOI	Total
Lower Riphean	215	16.47	0.11	2.46	2.00	0.07	12.49	28.73	0.26	0.91	0.07	179	290	36.19	99.76
Middle Riphean	262	11.11	0.08	1.91	1.55	0.07	10.23	35.26	0.16	0.39	0.04	161	266	38.43	99.23
Upper Riphean	558	7.95	0.09	1.66	1.37	0.06	7.38	40.91	0.18	0.38	0.04	111	447	39.10	99.12
Vendian	134	9.41	0.07	1.73	0.95	0.21	9.64	38.74	0.27	0.29	0.13	129	379	37.20	98.64

Notes: All oxides in wt %, Ba and Sr in ppm.

** The sum of FeO and Fe_2O_3.*

Upper Precambrian sequences, carbonates are dominated by dolomites (Qin *et al.*, 1985). Possible correlatives of the Upper Vendian deposits in the Sinian stratotype section in the Yangtze Gorges, the Doushanto and Dengying formations, are also represented largely by dolomites, with limestones clearly subordinate in significance (Chen, 1987). A wide distribution or the occasional predominance of limestones in carbonate rocks has been registered in the Upper Riphean in Spitsbergen, Norway, the Mackenzie Mountains and in the Victoria Lake area in Canada, and in the Adelaide "gleosyncline" in Australia (Knoll and Swett, 1990). This appears to imply that the discussed trend of changes in the Upper Precambrian carbonate rocks composition is a phenomenon going beyond one particular region in significance.

Comparison of data on the Upper Precambrian from the Urals and Siberia with data on the Phanerozoic rocks from the Russian and North American platforms (Vinogradov and Ronov, 1956; Ronov, 1970) shows that the general trend of changes of dolomite vs. limestones with time within the last 1.6 Ga was cyclic and directional, with the Late Riphean to the Early Paleozoic time characterized by a tendency toward an increasing relative role of dolomites. Mean compositions of carbonate rocks for more detailed stratigraphic units of the Phanerozoic (Vinogradov and Ronov, 1956) are strongly indicative of a more high-amplitude cyclicity of a lower order (Fig. 3). Whether this cyclicity reflects global processes or is typical only of the Russian platform remains to be understood.

Data on the compositions of carbonates from the Lower Proterozoic of the Baltic shield (Melezhik and Predovskii, 1989), if regarded as a reflection of global patterns of carbonate accumulation, gives reason to suggest that the Early Riphean maximum of dolomite formation was preceded by a Ludicovian (ca. 20 Ga) minimum, while in the Jatulian (2.3 - 2.0 Ga ago), the relative abundance of dolomites was even higher than in the Early Riphean of the studied regions.

Sr. Upper Riphean sections of the Urals and Siberia contain carbonate rocks relatively high in strontium (Fig. 4, Tables 1 and 2). Maximum Sr concentrations can be as high as 2000 ppm in the limestones of the Katav Formation of the Bashkirian anticlinorium, 5000 ppm in the Inzer Formation, or 3000 ppm in the Barakun and Kalancha formations in the southern Siberian platform. Sr-enriched carbonates in the Upper Riphean have also been recorded in the Polar and Subpolar Urals, Pai-Khoi, Timan (Yudovich *et al.*, 1980), the Patom Highlands and Igarka Uplift (Golovenok, 1985), in Greenland and Spitsbergen (Swett and Knoll, 1989), and in northern China (Qin *et al.*, 1985). This particular feature of the Upper Precambrian carbonates must also be more than regional in significance. Rocks high in strontium are usually limestones, often with an oncolite (pisolith) texture (Garrels, 1975; Yudovich *et al.*, 1980). Swett and Knoll (1989) have shown that the Sr content in oncolites of the rocks they studied were considerably higher than in the carbonate cement of the same rocks.

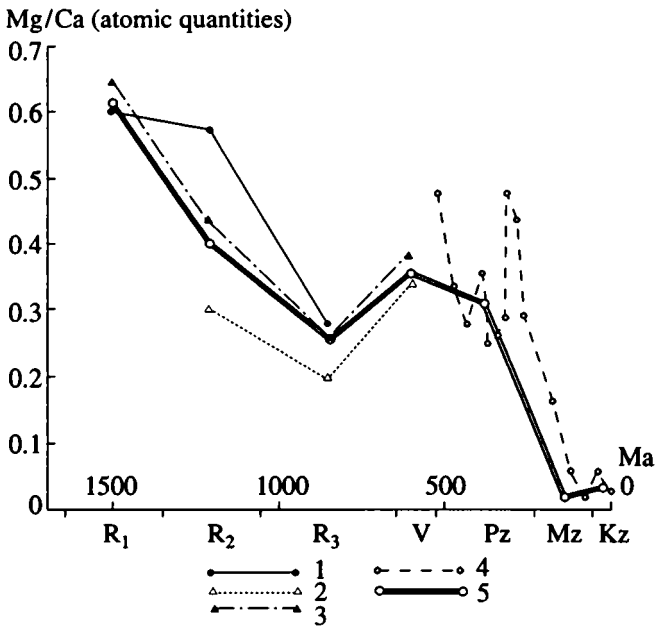

Fig. 3. *Variations in mean compositions of carbonate rocks throughout the Neogeikum. Data on the Phanerozoic from (Vinogradov and Ronov, 1956).*

1 – Urals; 2 – Siberian platform; 3 – folded fringes of the Siberian platform; 4 – Russian platform (mean compositions for relatively short time spans: periods, epochs); 5 – averaged over all regions discussed.

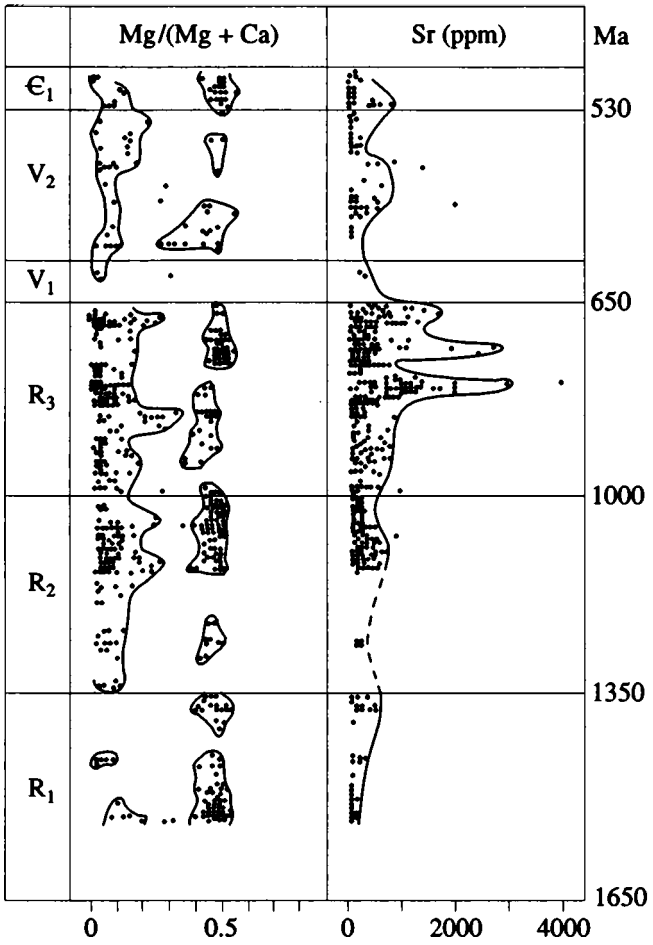

Fig. 4. *The value of modulus Mg/(Mg + Ca) (atomic quantities) and Sr abundances (in ppm) in Late Precambrian carbonate rocks. The plot combines data on the sections from the Bashkirian anticlinorium, Patom Highlands, and Uchur-Maya region.*

Because of similar ionic radii, strontium may replace calcium in the crystalline lattice of carbonate minerals. The capacity for Sr accumulation by means of such replacement progressively reduces almost by two orders of magnitude in the row of aragonite – high-magnesian calcite – low-magnesian calcite – dolomite, from the original 10,000 ppm, to a few hundred ppm (Carbonates ..., 1983; Yudovich *et al.*, 1988). With the diagenetic alteration of aragonite and high-magnesian calcite in low-magnesian calcite, Sr is expelled into the interstitial fluid of the sediments. In the open diagenetic system, given a free exchange of solutions across the sediment-water boundary, the residual concentrations of strontium in the emerging rocks would depend on the limits of its incorporation in lattices of low-magnesian calcites or dolomites. If, during sedimentation, the interstitial fluid somehow becomes isolated from the bottom waters, such fluid may precipitate Sr-bearing calcite or Sr minerals proper into the forming carbonates (Yudovich *et al.*, 1988). Thus, Sr concentrations in carbonates may depend on the original abundances of aragonite in the sediments and on the conditions of their diagenesis. High Sr concentrations in the Upper Riphean carbonates, discussed here, are observed in fairly thick and variable lithostratigraphic units, which occur within comparable stratigraphic ranges in different regions of the world. A phenomenon of such a scale would be difficult to account for with patterns of diagenesis. A more likely reason could be high Sr concentrations in the original composition of the sediments due to abundant aragonite or high-magnesian calcite in them. Because this pattern of composition of carbonate rocks distinguishes Upper Riphean deposits from the under- and overlying ones, the Earth must have known some periods in its history when, in contrast to Meso-Cenozoic and probably Late Riphean times, aragonite and high-magnesian calcite were not predominant synsedimentary carbonate minerals (Pigott and Mackenzie, 1979; Mackenzie and Pigott, 1981).

Mn. In the carbonate rocks that we studied, the highest (up to 3.2%) Mn content was fixed in the highly carbonaceous deposits at the base of the Vendian Tinnaya Formation in the south of the Siberian platform (Fig. 2), which can be interpreted as being a result of the anaerobic environment of sedimentation and the diagenesis of these deposits. Such carbonate and terrigenous rocks, which are rich in carbonaceous material, are widespread in the middle part of the Vendian sequences in the Siberian and Russian platforms and apparently are a manifestation of the euxinization of Late Vendian basins, which might have been global (Sochava, 1992). Elevated Mn concentrations have also been recorded in carbonate rocks of the Vendian Yasarusalin Formation of Vaigach Island and the Lower Riphean Korda Formation of the Enisei Ridge (Table 1).

Si and Al. The secular trend reflecting variations in mean concentrations of these elements has a positive correlation with the Mg trend (Table 2), illustrating, according petrographic data, variations in the amount of the terrigenous (mostly sandy) component in carbonate rocks. Its higher amounts could well be due to a more rugged topography of the source area or a higher saturation of coastal waters with carbonates. The proposed model of carbonate accumulation evolution in the Riphean-Vendian time allows for concurrent impact by these two factors.

EVOLUTION OF CARBONATE ACCUMULATION IN THE LIGHT OF A QUASI-STEADY MODEL OF THE GEOCHEMICAL CYCLE

When analyzing various factors responsible for the directional and cyclic changes in the composition of carbonate rocks throughout the Late Precambrian and Phanerozoic (Fig. 3), primary attention should be given to problems associated with the balances of carbon, calcium, and magnesium flows within the geochemical cycle (Fig. 5), and to possible variations of these values in time.

Calculations of the global rates of the accumulation of various chemical elements by sediments and a com-

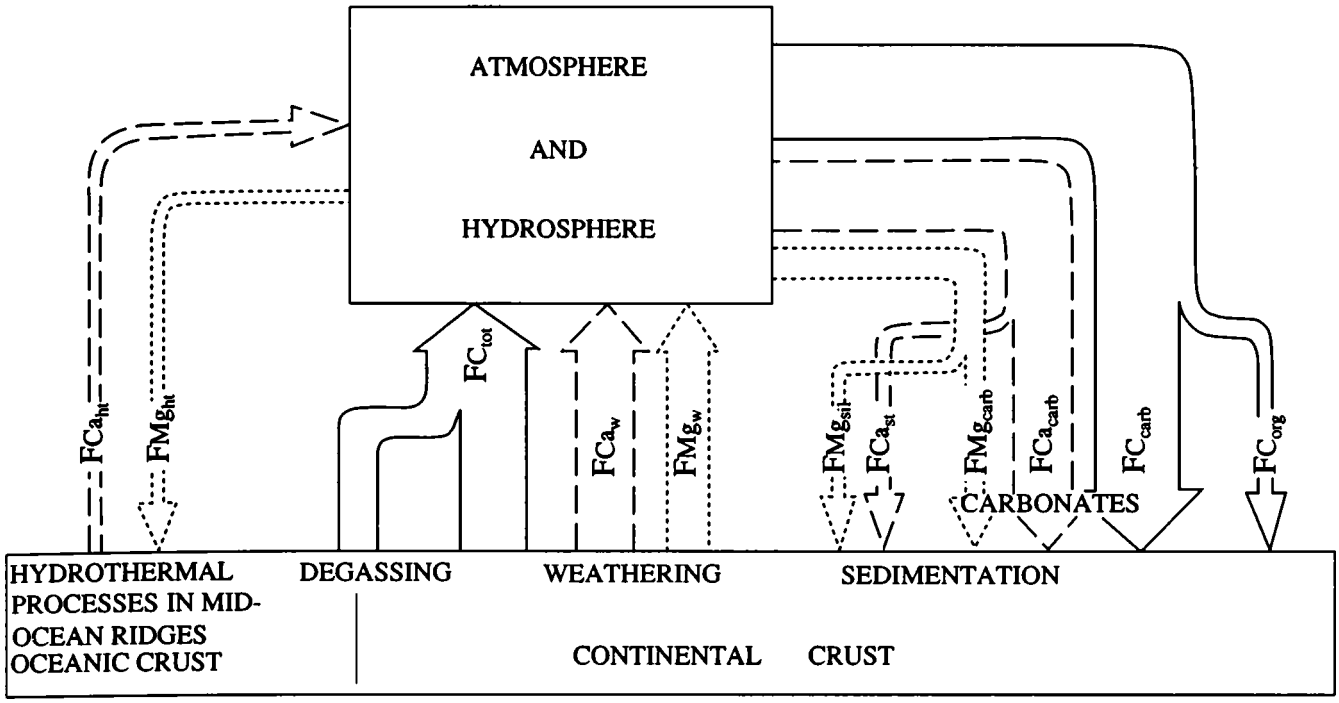

Fig. 5. *Geochemical cycle of carbon, calcium and magnesium.*
Flows: FC_{tot} is total C flow into atmosphere and hydrosphere due to degassing of solid Earth's shells and weathering in source areas; FCa_w and Fmg_w are flows of Ca^{2+} and Mg^{2+} in river discharges from areas of weathering on continents; FC_{carb}, FCa_{carb}, and FMg_{carb} are accumulations of carbon, calcium, and magnesium within carbonate sedimentary rocks; FC_{org} is accumulation of organic carbon in sediments; FCa_{st} and FMg_{sil} are accumulation in sediments of Ca in sulfates, and that of Mg in silicate (clay) rocks; FMg_{ht} and FCa_{ht} are Mg extraction from sea water and supply of Ca into sea water due to hydrothermal processes in the basalt layer of the oceanic crust.

parison of the data with abundances of these elements in the atmosphere and hydrosphere have led to models presupposing that the steady state for these spheres since the residence time of most of the elements within the atmosphere and hydrosphere system is much shorter than the life of the system (Garrels, 1975). Such a state of this complex system provides for equal rates of input and output flows of substances, as a result of various self-regulating mechanisms (Sochava, 1992). A few models, describing such mechanisms responsible for the quasi-steady state of the carbonate system of the ocean and atmosphere, follow below.

Regulation of the Mg/Ca ratio in sea water. One of the major parameters of the environment that define the degree of Mg involvement in carbonate accumulation is the Mg/Ca ratio in bottom waters. An increased value of this ratio would result in the modification of the mineral form of precipitating carbonates in the following sequence: low-magnesian calcite – high-magnesian calcite + aragonite – dolomite. With the general salinity decreasing, the threshold value of the Mg/Ca ratio that delimitates calcite-accumulating from dolomite-accumulating environments is brought down. As a result, in modern sediments, dolomite is much more widespread in fresh-water basins than in marine ones (Folk and Land, 1975; Carbonates ..., 1983; Kholodov, 1988).

In the most wide-spread magmatic rocks of the provenances and in mean estimations of the composition of the Late Precambrian provenances, magnesium and calcium abundances are commensurate (with the exception of andesites) with a certain but insignificant preponderance of magnesium (Table 3). In the mean compositions of sediments (of both siliciclastic and carbonate), calcium considerably dominates over magnesium. As a result of weathering and denudation in the provenances, most of the magnesium is supplied to sedimentary basins by river waters as an alumosilicate clayey-silty suspension, where Mg is much more abundant than Ca, as is suggested by the compositions of shales of different ages. For example, in the mean composition of the studied carbonate-free Upper Proterozoic shales, the Mg/Ca ratio is 4.36. As calcium reaches the sedimentation basins mostly as solution, the Mg/Ca ratio in the mean composition of modern river waters is noticeably lower than in the mean compositions of source-area rocks (Table 3). Although the influx of dissolved substances discharged now by rivers into the oceans has Ca^{2+} concentrations much higher than Mg^{2+}, the composition of the modern oceanic waters has a reverse pattern, with Mg/Ca of 4.4. Then, the estimated residence time of Ca^{2+} in oceanic waters (1.2 Ma) is by an order of magnitude shorter than that of Mg^{2+} (15 Ma) (Garrels, 1975).

Table 3. Abundances of MgO and CaO (wt %) and the Mg/Ca ratio (atomic quantities) in the most widespread types of rocks in source areas, the Earth's shells, and the hydrosphere

Rocks, Shells, etc.	MgO	CaO	Mg/Ca
(1) Granitoids and granite-gneisses	1.79	2.89	0.85
(2) Basalts of folded zones of continental crust	6.42	10.52	0.85
(3) Andesites of folded zones of continental crust	3.29	6.66	0.69
(4) A mean composition of shields	2.43	3.89	0.87
(5) Granite-metamorphic shell of continental crust	2.83	4.08	0.96
(6) Sedimentary shell (without effusives)	2.96	10.92	0.37
(7) Modern river water	0.41×10^{-3}	1.5×10^{-3}	0.38
(8) Marine water, salinity 35‰	0.129	0.041	4.38
(9) Upper Proterozoic Carbonate-free mudstones	2.47	0.78	4.36

Notes: Sources: 1 - 6 – Ronov et al., (1990); 7, 8 – Taylor and MacLennan (1985); 9 – the authors' data on 1912 analyses from the same sections as the carbonate rocks discussed in this paper.

Were marine and oceanic basins to accumulate solely calcium carbonates, then, as is dictated by the Mg/Ca ratio in the influx of substances presently supplied to this system, the Mg/Ca ratio in the sea water should be growing, with only Ca^{2+} being removed and, when a certain threshold value is attained, Mg carbonate precipitating should be triggered. Carbonate accumulation in the form of dolomites, i.e., with a Mg/Ca ratio much higher than in the input substances, must reduce the Mg/Ca ratio in the sea water and ultimately suppress the accumulation of Mg carbonates. This negative feedback between the Mg/Ca ratio in the sea water and the rate of accumulation of dolomites in the sediments maintains such a value of the Mg/Ca ratio, which is required for balanced Mg^{2+} and Ca^{2+} flows at the input and output of the ocean's system.

Regulation of pCO_2 and temperature at the surface of the Earth. The rate of input of CO_2 in the atmosphere-hydrosphere system varied in the geologic history, thus affecting not only the overall carbonate accumulation rate (Ronov, 1976), but their mineral composition as well. The following agents are thought to affect the composition of carbonate sediments because of recurring variations in the CO_2 input rate.

(1) The composition of precipitating carbonate may vary because of the changes in pCO_2, with the Mg/Ca ratio in the sedimentation environment remaining constant. A low pCO_2, like today's, would favor deposition of high-magnesian calcite and aragonite, whereas a higher pCO_2 would be more favorable to low-magnesian calcite and dolomites (Pigott and Mackenzie, 1979; Mackenzie and Pigott, 1981). Such an effect was already inferred by Strakhov (1951) to explain the evolution of carbonate compositions in the Earth's history.

(2) The increasing of Mg content in the carbonate sediments can also be due to a higher temperature in the environment, related to the greenhouse effect of high CO_2 concentrations in the atmosphere. This was thoroughly examined for biogenic carbonates. The most sensitive organisms affected by this factor are red algae, in whose carbonate skeletons $MgCO_3$ concentrations may vary from 5 to 30 mol %, within the temperature range of 30°C (Carbonates ... , 1983).

(3) According to the model proposed by Walker *et al.*, (1981), the growth of pCO_2 and surface temperatures should boost the erosion of continents and add to the input of Ca-group elements into the oceans and, consequently, to the carbonate accumulation rate. Note that calcium seems to be thoroughly removed from the source rocks at an average stage of their weathering. With crusts of weathering expanding on the continents, where the degree of primary rock alteration is close to that of the kaolinite stage, the input of Ca-group elements would grow mostly as a result of a higher Mg contribution. The Mg/Ca ratio in this inflow should grow to nearly that of the source rocks (Table 3). This would ultimately enhance the role of dolomite in the marine sediments.

The above mechanisms, employed by the atmosphere-hydrosphere carbonate system in response to faster CO_2 input rates, favor higher involvement of the sea water Mg in the process of carbonate accumulation. Because the majority of the dissolved Ca input into the system precipitates into carbonate rocks (see below), the appearance of an additional cation in the carbonate accumulation would mean a greater volume of the deposited carbonate rocks and, hence, higher rates of CO_2 removal from the system. Thus, the self-regulating mechanism responsible for the relative stability of pCO_2 in the atmosphere and hydrosphere and of the surface temperatures involves not only variations in the intensity of weathering in the source areas (Walker *et al.*, 1981), but also variations in the general rate of dolomite accumulation in the sediments, which may be compensated by the inversely variable rates of Mg^{2+} removal from the oceanic waters, due to other agents.

Regulation of Mg^{2+} consumption by hydrothermal processes in the oceanic crust. The discovery and the studies of the phenomenon of sea water circulation in hydrothermal systems of mid-oceanic ridges encouraged some researchers to review the exclusive role of river discharge and sedimentation for the specific chemical and isotopic compositions of oceanic water. This process is believed to remove Mg^{2+} from the oceans in amounts comparable to the river input. Another effect of this process is the high influx of calcium in sea water (Holland, 1984).

Secular variations of the rate of Mg^{2+} removal from the oceanic water due to the above process can be viewed as a function of the changing rate of the spreading and the overall length of mid-oceanic ridges, i.e., as a phenomenon independent of the state of the ocean's carbonate system, but instead, to a large extent, affecting it. On the other hand, the amount of Mg^{2+} consumed by high-temperature reactions involving basalts may also be dependent on its concentration in the sea water: in a way, it is a passive channel to dump "excess" magnesium that has not been consumed by carbonate-accumulation processes. Similarly, the rate of accumulation of authigenic magnesium in the silicate phase of marine sediments (in clays) seems also to be defined by its concentration in sea water. Then, it must be an elementary case of negative feedback: higher concentrations of an element lead to faster rates of its removal from the system.

The balance of calcium, magnesium and carbon flows and the evolution of carbonate accumulation. Based on the assumption of the quasi-steady state of the atmosphere-hydrosphere system, the balance of carbon flows between the outer spheres of the Earth can be represented in the equation:

$$FC_{tot} = FC_{carb} + FC_{org}. \quad (1)$$

For a designation of the flows, see Fig. 5. The flow FC_{tot} comprises carbon dioxide and other carbon compounds that are supplied as part of the products of the crust and mantle degassing by volcanic and metamorphic processes subsequently oxidized to CO_2, and the carbon supplied into the hydrosphere in the form of HCO_3^- ions as a result of the weathering of carbonate and organic carbon-bearing rocks in the source areas.

The relative consistency of the isotopic composition of carbon in marine sedimentary carbonates from the Early Precambrian through the Cenozoic (Schidlowski, 1982) can be regarded as evidence of a consistency in time of a ratio of: $FC_{org}/FC_{tot} \approx 0.2$, or

$$FC_{carb} \approx 0.8FC_{tot}. \quad (2)$$

The dissolved calcium is supplied to the sea water as a result of the weathering of carbonite, sulfate, and silicate rocks in the continental source areas. Another oceanic-calcium source must be the interaction of sea water with basalts under hydrothermal processes in mid-oceanic ridges (Holland, 1984). The majority of the calcium goes to sediments to form carbonate rocks, while a minor portion makes up sulfates (Fig. 5). Hence,

$$FCa_{carb} = k(FCa_w + FCa_{ht}), \quad (3)$$

where k is a coefficient reflecting the time-variable proportion of carbonate Ca in the overall flow of calcium removed from the oceanic water into the sediment. According to Garrels's estimates of Phanerozoic means of element flows from oceans to sediments (Garrels, 1975), calcium precipitated as sulfates accounts for 6% of the total Ca^{2+} flow, or a meagre 4%, according to Ronov *et al.*, (1990). Phosphates and other authigenic Ca minerals known in the sediments can be neglected in evaluating the general balance. Hence, the mean value of $k \approx 0.95$, i.e., fairly close to one.

The dissolved magnesium is supplied to the sea water as a result of weathering and erosion of the source-area, to be consumed by carbonate accumulation, authigenic accumulation or alteration of clay minerals, and by hydrothermal processes in the oceanic crust (Fig. 5), or

$$FMg_w = FMg_{carb} + FMg_{sil} + FMg_{ht}. \quad (4)$$

The respective roles of the two latter channels of Mg^{2+} discharge from the oceans are as yet uncertain.

The summary value of FCa_{carb} and FMg_{carb} is controlled by the rate of carbon input into the system:

$$FCa_{carb} + FMg_{carb} \approx 0.8FC_{tot}. \quad (5)$$

The majority of Ca^{2+} accumulates in the sediments as constituents of the carbonates, while Mg^{2+} has other important channels to escape the oceanic waters. The changes in the atmosphere-hydrosphere system required for the deposition of dolomites are brought about by growing volumes of the reservoirs and, consequently, by the growing residence time in the system of the primary dolomite-forming components: CO_2 and Mg^{2+}. Analysis of the self-regulatory mechanism in the system suggests that Ca^{2+} is a cation with a certain priority with respect to Mg^{2+}, when involved in the process responsible for the passage of CO_2 and its derivatives from the atmosphere and hydrosphere to carbonate sediments. In other words, Mg^{2+} is involved in carbonate accumulation solely in so far as the carbon flow that passes through the atmosphere-hydrosphere system exceeds in volume the calcium flow and hence, cannot be buried in the sedimentary rocks without an additional cation. Based on these prerequisites and on equations (2), (3), and (5), the interplay between the flows of Mg, Ca, and C in the course of carbonate accumulation can be represented as:

$$FMg_{carb} \approx 0.8FC_{tot} - k(FCa_w + FCa_{ht}), \quad (6)$$

while the evolution of the composition of carbonate rocks as a consequence of the variation of values of these flows with time, or

$$dFMg_{carb} \approx 0.8dFC_{tot} - k(dFCa_w + dFCa_{ht}). \quad (7)$$

Secular variations of the values of the three flows in the right-hand part of the equation may be due to the variable intensity of the geodynamic and magmatic processes. Stronger processes should entail higher flow rates. However, the rate increase is not adequate. The range of variations of the input rates for elements supplied into the atmosphere-hydrosphere system by weathering (Fig. 5) is lower than the range of variations of the flows caused by degassing, as suggested by cyclic variations in the proportion of the volatiles, carbon in particular, in the composition of sedimentary rocks from the different Phanerozoic series, or

$$0.8|dFC_{tot}| > k|dFCa_w|. \quad (8)$$

Relations between variations of FC_{tot} and FCa_{ht} can be assessed using geological data by comparing

changes with time in the general composition of carbonate rocks and the tectono-magmatic activities. If

$$0.8|dFC_{tot}| > k|(dFCa_w + dFCa_{ht})|, \quad (9)$$

epochs of high tectono-magmatic activity should result in a higher dolomite proportion in the carbonate rocks, whereas limestones would dominate in the opposite cases. With the Late Riphean and Vendian-Paleozoic stages in the Earth's history as examples, the conclusion will be that the former assumption seems to be more consistent with the geohistorical data. The Pangaea-I supercontinent that had existed in the Riphean time was destroyed at the Late Riphean-Vendian boundary, giving way to new oceans, the Proto-Tethys in the Late Riphean, and the Central Asian and Yapetus in the Vendian (Borukaev, 1985; Khain and Bozhko, 1988), and resulting in intense volcanism, accompanied by degassing of the crust and mantle (higher FC_{tot} flow) and, apparently, by higher hydrothermal activity in the newly-formed mid-oceanic ridges (higher FCa_{ht} flow). One of the criteria of the growing intensity of spreading has traditionally been regarded as eustatic sea-level rises due to larger volumes of mid-oceanic ridges (e.g., Holland, 1984). The accumulation of the Upper Vendian deposits at the base of sedimentary covers on the Russian, Siberian, and other cratons took place under conditions of wide transgression of the shelf seas that, at the time, covered much larger areas of the continental crust than did the Late Riphean sedimentary basins (Semikhatov, 1974; Khain and Bozhko, 1988). It follows, then, that dolomites were becoming relatively more prominent from the Late Riphean to the Vendian and Early Paleozoic, in the background of higher tectono-magmatic activity (Fig. 3), and this growth was definitely associated with higher rates of CO_2 input in the atmosphere and hydrosphere at that time span.

The above model has the virtue of providing a geologically consistent explanation of cyclic variations in the dolomite/limestone ratio in carbonate sequences throughout the geologic history. At the same time, the model seems unlikely to be applicable to the explanation of the main trend in the evolution of carbonates, viz., the lowering of the above ratio. A progressive decline of the rates of CO_2 input in the atmosphere and hydrosphere throughout history in the quasi-steady state of the system must have resulted in a decrease in overall volume of the deposited carbonates, which contradicts the available data (Ronov *et al.*, 1990), pointing rather to the growth of this parameter of global sedimentation with time. Thus, the reason underlying this long-term trend must be the growing Ca inflow into the oceanic basins.

An exchange in flows between the atmosphere, hydrosphere, and the sedimentary shell is executed on the principle of open dynamic systems, and, with the quasi-steady state attained, no changes in these shells occur because of the equivalence of input and output flows. The interaction between the hydrosphere and the basaltic layer of the oceanic crust is different. A probable effect of hydrothermal processes in oceanic basalts is a unidirectional inflow of calcium into the hydrosphere-sedimentary shell system (and its loss of Mg); as a result, Ca amounts in the sedimentary shell should increase with time, whereas those of magnesium should decrease. This model is supported by the evidence of excess calcium in the sedimentary shell (Ronov *et al.*, 1990), which is twice its average concentration in the continental crust (Table 3). Because sedimentary rocks in the Neogeikum represented the majority or a major proportion of the rocks exposed to weathering and denudation in the source area, it is logical to expect a gradual increase in the rates of the FCa_w flow.

CONCLUSION

The mechanisms of self-regulation at work in the Earth's atmosphere and hydrosphere system are capable of setting values of major carbonate-accumulation parameters (pCO_2, surface temperature, Mg/Ca ratio in sea water, etc.), which are necessary to maintain the quasi-steady state of this system, which implies the equivalence of Ca, Mg, and C input and output. Because of these mechanisms, the global rates of carbonate accumulation is limited to the input of CO_2 in the atmosphere-hydrosphere system; Mg^{2+} is to be involved in carbonate accumulation only to the extent by which the CO_2 inflow in this system exceeds the Ca^{2+} flow. According to this model, any decline in the dolomite/limestone ratio in carbonates of this or another epoch can be due either to the growing input of Ca^{2+} or to the reduced input of CO_2 in the atmosphere and hydrosphere.

The enrichment of sea water in Ca^{2+} by hydrothermal processes in the oceanic crust must be responsible for the higher concentrations of this element in the sedimentary shell. It is apparently because of this irreversible process that the Earth's sedimentary shell is twice as rich in calcium as the continental crust on average (Ronov *et al.*, 1990). Because of the rate increases of Ca_w input into the oceanic waters as it accumulated in the sedimentary shell in the course of geologic history, the relative role of limestones became progressively more important, with a respective decline of dolomites in the global carbonate accumulation.

Our data on the secular variations in the composition of carbonate rocks in the Late Precambrian add to the concepts of cyclic and directional mode of the evolution of carbonate accumulation. Throughout the Neogeikum, two long-term cycles characterizing variations in the dolomite/limestone ratio can be delineated, with minima of this ratio in the Late Riphean and Meso-Cenozoic. The above cycles are superimposed on the general trend toward the reduction of the ratio (Fig. 3). Both Meso-Cenozoic (Vinogradov and Ronov, 1956; Holland, 1989) and Late Riphean carbonates are high in strontium, which may point to the wide extent of aragonite in the original composition of sediments of this age.

In terms of the proposed model of the evolution of carbonate accumulation, the minimum of dolomite formation registered in the Late Riphean must have resulted from lower rates of CO_2 input in the atmosphere due to the global decline of tectono-magmatic processes (Borukaev, 1985). A reactivation of these processes in the Late Vendian and Early Paleozoic led to more intense dolomite formation. Such interpretation of the evolution of exogenic conditions in the late Precambrian is consistent with the broad expansion of the Late Riphean and Early Vendian continental glaciation (Chumakov, 1978), which can be regarded as a consequence of the weakening of the greenhouse effects as CO_2 concentrations in the atmosphere lowered; it is also consistent with earlier interpretations of data on the variation of sulfur isotopes in sulfates and of carbon and strontium isotopes in carbonates, registered in that segment of geologic history (Sochava, 1992).

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REFERENCES

Borukaev, Ch.B., *Struktura Dokembriya i Tektonika Plit* (Precambrian Structure and Plate Tectonics), Nauka, Novosibirsk, 1985.

Carbonate Rocks. Origin, Occurrences, and Classification, *Develop. Sedimentol.*, vol. 9A, Elsevier, Amsterdam, Oxford, New York, 1967a.

Carbonate Rocks. Physical and Chemical Aspects, *Develop. Sedimentol.*, vol. 9B, Elsevier, Amsterdam, Oxford, New York, 1967b.

Carbonates: Mineralogy and Chemistry, *Rev. Mineral.*, 1983, vol. 11.

Chen Ping, *The Sinian System, Stratigraphic Excursion Guidebook in the Yangtze Gorges Area*, Beijing, 1987, pp. 2 - 7.

Chumakov, N.M., *Dokembriiskie Tillity i Tilloidy* (Precambrian Tillites and Tilloids), Nauka, Moscow, 1978.

Daly, R.A., First Calcareous Fossils and the Evolution of the Limestones, *Bull. Geol. Soc. Amer.*, 1909, vol. 20, pp. 153 - 170.

Folk, R.L. and Land, L.S., Mg/Ca Ratio and Salinity: Two Controls Over Crystallization of Dolomite, *Bull. Amer. Assoc. Petrol. Geol.*, 1975, vol. 59, no. 1, pp. 60 - 68.

Garrels, R.M., *Krugovorot Ugleroda, Kisloroda i Sery v Techenie Geologicheskogo Vremeni* (Circulation of Carbon, Oxygen, and Sulfur throughout the Geological Time), Nauka, Moscow, 1975.

Golovenok, V.K., On Abnormally High Strontium Abundances in the Riphean Onkolite Limestones of the Patom Highlands, *Litol. Polezn. Iskop.*, 1985, no. 1, pp. 122 - 127.

Holland, H., The Chemical Evolution of Atmosphere and Oceans, Princeton University Press, 1984.

Khain, V.E. and Bozhko, N.A., *Istoricheskaya Geotektonika* (Historic Geotectonics), Nedra, Moscow, 1988.

Kholodov, V.N., Problems of Dolomite Formation at the Present Stage of Lithology, *Evolutsiya Karbonatonakopleniya v Istorii Zemli*, Nauka, Moscow, 1988, pp. 3 - 23.

Khomentovsky, V.V., Shenfil', V.Yu., and Yakshin, M.S., Baikalian of Siberia and the General Late Precambrian Scale, *Stratigrafiya Verkhnego Proterozoya SSSR (Rifei i Vend), Tez. Dokl., 2 Vsesoyuzn. Soveshch. "Obshchie Voprosy Raschleneniya Dokembriya SSSR"* (2nd All-Union Conf. "General Problems of the USSR Precambrian Subdivision"), Ufa, 1990, pp. 18 - 19.

Knoll, A.H. and Swett, K., Carbonate Deposition during the Late Proterozoic Era: an Example from Spitsbergen, *Amer. J. Sci.*, 1990, vol. 290-A, pp. 104 - 132.

Mackenzie, F.T. and Piggot, J.D., Tectonic Controls of Phanerozoic Sedimentary Rock Cycling, *J. Geol. Soc. London*, 1981, vol. 138, part 2, pp. 183 - 196.

Melezhik, V.A. and Predovskii, A.A., Karelian Inversion of Carbonate Accumulation on the Baltic Shield: Reflection of Global Variations in Precambrian Exogenesis?, *Dokl. Akad. Nauk SSSR*, 1989, vol. 306, no. 6, pp. 1441 - 1445.

Petrov, V.G., *Dokembrii Zapadnogo Obramleniya Sibirskoi Platformy (Geologiya i Petrokhimiya)* (Precambrian of the Western Fringes of the Siberian Platform (Geology and Petrochemistry)), Nauka, Novosibirsk, 1982.

Piggot, J.D. and Mackenzie, F.T., Phanerozoic Ooid Diagenesis: A Signature of Paleo-Ocean and Atmospheric Chemistry, *Abstr. with Programs, Bull. Geol. Soc. Amer.*, 1979, vol. 11, no. 7, pp. 495 - 496.

Qin Zhengyuong, Yang Xiuen, and Jiang Mingmei, Chemostratigraphic Correlation of the Middle and Upper Proterozoic between the Yanshan and Shennongjia Basins, *Precambr. Res.*, 1985, vol. 29, pp. 77 - 91.

Ronov, A.B., Chemical Composition and Conditions of Formation of Paleozoic Carbonate Sequences of the Russian Platform (from Data of Litho-Geochemical Maps), *Tipy Dolomitovykh Porod i Ikh Genezis*, Nauka, Moscow, 1956, pp. 256 - 343.

Ronov, A.B., Volcanism, Carbonate Accumulation, Life (Regularities of Carbon's Global Geochemistry), *Geokhimiya*, 1976, no. 8, pp. 1252 - 1277.

Ronov, A.B. and Migdisov, A.A., Evolution of Chemical Composition of Rocks of Shields and the Sedimentary Cover of the Russian and North American Platforms, *Geokhimiya*, 1970, no. 4, pp. 403 - 438.

Ronov, A.B., Yaroshevskii, A.A., and Migdisov, A.A., *Khimicheskoe Stroenie Zemnoi Kory i Geokhimicheskii Balans Glavnykh Elementov* (Chemical Structure of the Earth's Crust and Geochemical Balance of Major Elements), Nauka, Moscow, 1990.

Schidlowski, M., Content and Isotopic Composition of Reduced Carbon in Sediments, *Mineral Deposits and Evolution of the Biosphere*, Springer, Berlin, 1982, pp. 103 - 122.

Semikhatov, M.A., *Stratigrafiya i Geokhronologiya Proterozoya* (Stratigraphy and Geochronology of the Proterozoic), Nauka, Moscow, 1974.

Semikhatov, M.A. and Serebryakov, S.N., *Sibirskii Gipostratotip Rifeya* (Siberian Hypostratotype of the Riphean), Nauka, Moscow, 1983.

Sochava, A.V., Quasi-Stationary Model of the Geochemical Cycle and Biosphere's Evolution at the Precambrian-Phanerozoic Boundary, *Izv. Ross. Akad. Nauk, Ser. Geol.*, 1992, no. 6, pp. 41 - 56.

Strakhov, N.M., *Izvestkovo-Dolomitovye Fatsii Sovremennykh i Drevnikh Vodoemov* (Calcareous-Dolomitic Facies of Modern and Ancient Basins), Akad. Nauk SSSR, Moscow, 1951.

Stratotip Rifeya. Stratigrafiya i Geokhronologiya (Riphean Stratotype. Stratigraphy and Geochronology), Nauka, Moscow, 1983.

Swett, K. and Knoll, A.H., Marine Pisolites from Upper Proterozoic carbonates of East Greenland and Spitsbergen, *Sedimentol.*, 1989, vol. 36, pp. 75 - 93.

Taylor, S.R. and MacLennan, S.M., The Continental Crust, Its Composition and Evolution, Canberra, 1985.

Veizer, J., Secular Variations in the Composition of Sedimentary Carbonate Rocks, I. Alkali Metals, *Precambr. Res.*, 1978, vol. 6, pp. 367 - 380.

Veizer, J., Secular Variations in the Composition of Sedimentary Carbonate Rocks, II. Fe, Mn, Ca, Mg, Si, and Minor Constituent, *Precambr. Res.*, 1978, vol. 6, pp. 381 - 413.

Vinogradov, A.P. and Ronov, A.B., Composition of Sedimentary Rocks of the Russian Platform in Connection with Its History of Tectonic Motions, *Geokhimiya*, 1956, no. 6, pp. 533 - 559.

Walker, J.C.G., Hays, P.B., and Kasting, J.F., A Negative Feedback Mechanism for the Long-Term Stabilization of Earth's Surface Temperature, *J. Geophys. Res.*, 1981, vol. 86, no. C10, pp. 9776 - 9782.

Yudovich, Ya.E., Maidl', T.V., and Ivanova, T.I., *Geokhimiya Strontsiya v Karbonatnykh Otlozheniyakh* (Strontium Geochemistry in Carbonate Deposits), Nauka, Leningrad, 1980.

Yudovich, Ya.E., Tereshko, V.V., Gareev, E.Z., and Oparenkova, L.I., Strontium Geochemical Horizons in Riphean Carbonate Deposits of the Eastern European Platform and the Northern Urals, *Trans. Inst. Geol., Komi Scientific Center, Urals Branch of the USSR Acad. Sci.*, Syktyvkar, 1988, vol. 67, pp. 90 - 106.

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Evolution of the biosphere

The First Finds of Organic-Walled Microfossils in the Upper Precambrian of the Bokson-Sarkhoi Depression (the Eastern Sayan)

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Abstract – Organic-walled microfossils of the Dibi Formation and the Sarkhoi and Bokson groups, widely distributed in the southeast part of the Eastern Sayan, were studied. No microfossils had been found *in situ* until this time in the above units. Previously published references on the occurrence of a complex Paleozoic acritarch in these units were disavowed by the decisions of the Fifth Colloquium on Microfossils in Leningrad in 1986. The first microphytological characteristics of the Dibi (the middle part), the upper Sarkhoi (Khushatai), and the lower Bokson (Tabinzurta) formations are presented. The most common transit Late Precambrian taxa, such as acritarchs *Leiosphaeridia* and filamentous *Leiotrichoides* and/or *Eomycetopsis* (the whole studied sample includes seven genera and 10 species), dominate in all three formations. Only the upper Sarkhoi (Khushatai) deposits contain more distinct *Polytrichoides lineatus* and *Granomarginata judomica*, which allow clarifying the stratigraphic age of the fossiliferous unit and assigning the Khushatai Formation of the Sarkhoi Group to the Vendian.

The age and the relationships of a number of lithostratigraphical units in the southeast part of the Eastern Sayan (the Bokson-Sarkhoi depression) have been subjects of lively discussion in recent publications. According to traditional views, the carbonate-sandy-shaly Dibi Formation (up to 1000 m) and the volcanoterrigenous Sarkhoi (up to 2500 m) and the carbonate Bokson (up to 1000 m) groups are found in a normal stratigraphic succession and are assigned to the upper horizons of the Riphean, to the Vendian and the Early Cambrian (Semikhatov and Serebryakov, 1967; Zhabin and Stepanova, 1974; Dobretsov, 1985; Khomentovskii *et al.*, 1985; Kuz'michev, 1990). However, some authors attribute the first two of these units to the Early or Middle Paleozoic, considering them to form individual tectonic nappes. This viewpoint reflects the existence of objectively identified large thrusts in the region (Dobretsov, 1985) and the scantiness and contradictions of the published paleontological data on the age of the above-listed units. The archaeocyates of the Early Cambrian Bazaikha and Kameshki horizons, discovered approximately in the middle part of the Bokson Group (in the lower part of the Khuzhirtai Formation), are, essentially, the only reliable biostratigraphic markers in the region. These archaeocyates, along with stromatolites reported in the lower part of the Bokson Group, references on the occurrence of medusoids and the most ancient small shelly fossils found here, and the recent whole rock Rb-Sr dating of the Sarkhoi volcanics, equal to 718 ± 30 Ma (Buyakaite *et al.*, 1989), are a base for drawing the lower boundary of the Cambrian in the lower part of the Bokson Group. At the same time, the references on the occurrence of the Lower and Middle Paleozoic algae, bryozoans, shell fragments, and microfossils in the regarded strata were published recently (see Butov, 1985, 1986, for bibliography). In particular, broad sets of organic-walled microfossils, including complex acanthomorphic acritarchs known in Cambrian-Devonian deposits of other regions, were mentioned in Dibi, Sarkhoi, and Bokson deposits. Neglecting the other mentioned fossils, we shall note that the Eastern Sayan preparations thought to contain the cited microfossils were examined at the Fifth All-Union Colloquium on Precambrian Microfossils in Leningrad (Yankauskas *et al.*, 1987). This examination revealed a complete absence not only of morphologically differentiated Paleozoic forms in them, but of any recognizable microfossils at all. Thus, there is no objective information about Dibi-Bokson microfossils in the published papers. The data presented below on the microfossils from the Dibi, Khushatai, and Tabinzurta formations of the Bokson-Sarkhoi depression give such information for the first time.

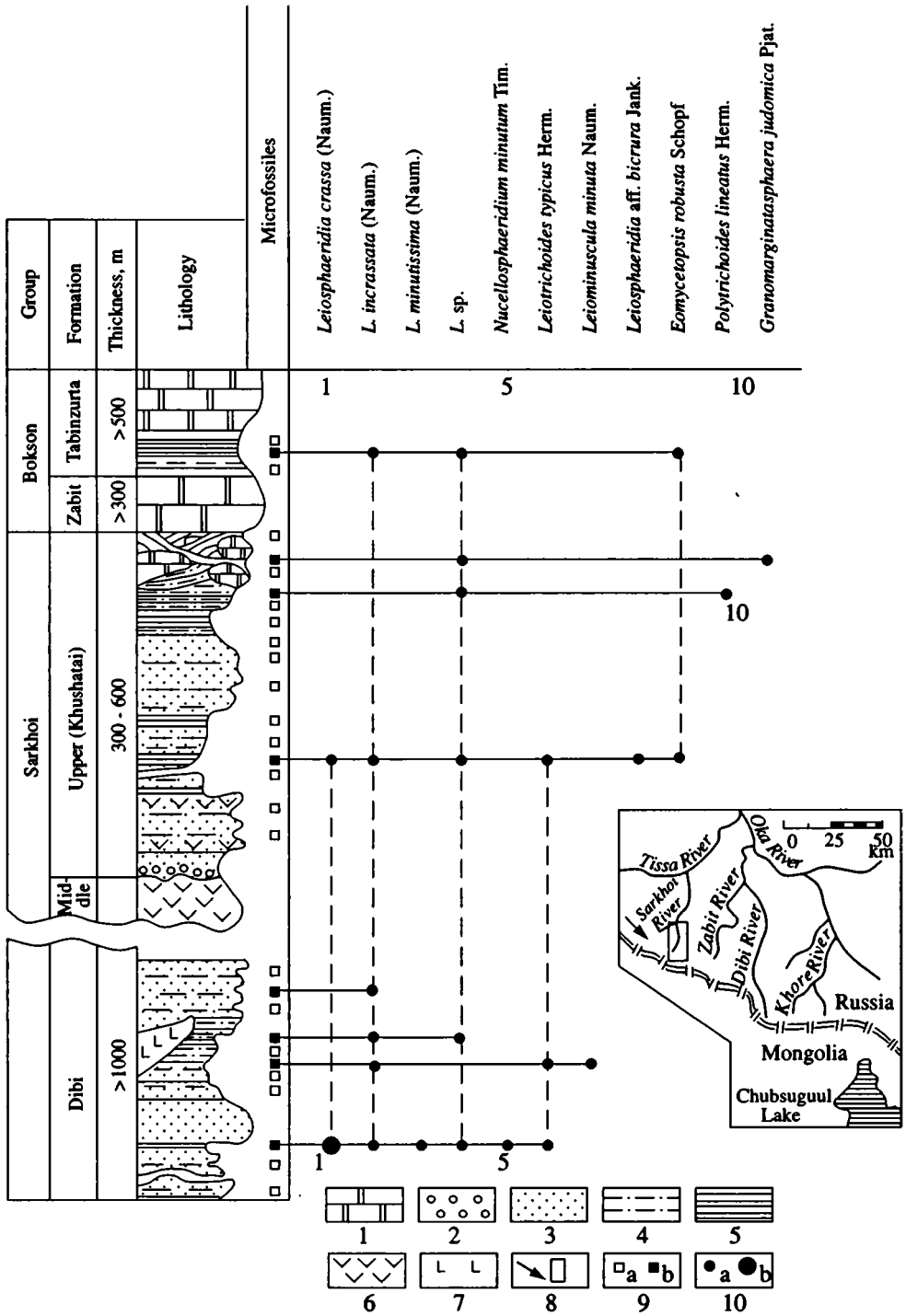

Fig. 1. *Distribution of microfossils in the Upper Precambrian deposits of the Bokson-Sarkhoi depression (upper reaches of the Sarkhoi River).*

1 – dolomites; 2 – conglomerates; 3 – sandstones; 4 – siltstones; 5 – shales; 6 – volcanics; 7 – diabase dikes; 8 – location of the studied region; 9 – position of collected samples: a – without microfossils, b – with microfossils; 10 – quantitative estimation of microfossils located at particular stratigraphic level: a – single specimens, b – dozens of speciments.

The studied material was collected in 1990 from a basin of the upper reaches of the Sarkhoi River, from the following units: (1) the Dibi Formation, composed of thin-rhythmical interbedding of siltstone-sandy and subordinate carbonate members; (2) the Sarkhoi Group, represented by an alternation of various silicoclastic and volcanic rocks (from its upper, Khushatai Formation); (3) the Tabinzurta Formation of the Bokson Group, dominated by phytolitic and silicified dolomites with subordinated interlayers of silicoclastic and clayey rocks. Tectonic contacts of these units within the upper reaches of the studied Sarkhoi River area open up the possibility for various interpretations of their stratigraphic succession (Resheniya ... , 1983; Butov, 1985, 1986; Kuz'michev, 1990). For microphytological prospecting, we used natural exposures of all perspective rocks of the listed stratigraphic units along the valleys of the creeks feeding into the Sarkhoi and Malaya Kharaganta rivers, in the area from the outfall of the Malaya Kharaganta to the upper reaches of the Sarkhoi (Fig. 1). Seventy-two samples were treated by a full laboratory maceration, twenty of them revealing microfossils.

Sampling of the Dibi Formation took place in the canyon-shaped valley of a small, nameless creek feeding into Urdo-Khundy Creek (right tributary of the Sarkhoi River), 3.5 - 4 km above its mouth. The samples originate from a 160- to 180-meter-thick section of the middle part of the formation, where gray and dark gray siliceous siltstones and shales alternate rhythmically with gray, mostly fine-grained sandstones. Seven out of 16 collected samples reveal definable microfossils. They are few in number (a few dozen specimens), mostly simple acritarchs with a diameter of 10 - 70 μm (more than 70% of all of them fall within the 20 - 30 μm diameter range). Present among them were *Leiosphaeridia crassa* (Naum.), *L. incrassata* (Naum.), *L. minutissima* Naum., *L.* sp., *Nucellosphaeridium minutum* Tim., very fine (less than 10 μm) *Leiominuscula minuta* Naum., and very rare sheaths of *Leiotrichoides typicus* Hermann (7 - 8 μm wide). The microfossils are satisfactorily preserved and, though most of them were corroded intensively, are yellowish gray, which testifies to the insignificant degree of alternation of the enclosing rocks.

The Sarkhoi Group samples were collected in a number of sites located in lower and upper parts of the section. Microfossils were absent in all 12 samples from siltstones interbedded with cross-laminated, mostly red-colored sandstones and gravelstones, composing a 100-meter-interval of the Nizhnii Sarkhoi Formation section (upper reaches of Munku-Sasan Creek). We also did not find any microfossils in 16 samples from the representative section (about 300 m thick) of the upper Sarkhoi Khushatai Formation in the Sokhor-Zhelga Creek valley (the left tributary of the Sarkhoi River). Rare microfossils were discovered only in the uppermost part of the section in four out of nine samples collected in a 10- to 12-m-thick member of green siltstones and shales, which crops out directly near the tectonic contact of the Sarkhoi Group and, supposedly, the lower Bokson Zabit Formation. Intensively corroded acritarchs *Leiosphaeridia* sp. (20 - 25 μm in diameter) and typical unsheathed multitrichomic filamentous *Polytrichoides lineatus* Hermann were found in the lower part of this member, and acritarchs *Leiosphaeridia* sp. (20 μm in diameter) and original volumetric thick-walled vesicles, which could be assigned to *Granomarginatasphaera judomica* Pjat. (14 - 15 μm in diameter), were discovered 6 - 8 m higher in the section. In addition to the above-noted locality, the upper Sarkhoi microfossils were discovered in one more point, some 150 - 170 m stratigraphically below the member containing the above-considered microfossils, in a 10-m-thick unit of gray siltstones and shales exposed in the upper reaches of the right nameless tributary of the Sarkhoi River, joining it 1.5 - 2 km above the mouth of Saganbaisu Creek. We found here the single acritarchs *Leiosphaeridia crassa* (Naum.), *L. incrassata* (Naum.), *L.* sp., and *L.* aff. *L. bicrura* Jank. (diameter 20 - 40 μm) and the sheaths *Eomycetopsis robusta* Schopf and *Leiotrichoides typicus* Hermann (3 - 8 μm wide) in four out of eight macerated samples.

Prospecting of the Bokson Group rocks was carried out in the numerous points situated mainly on the left bank of the upper reaches of the Sarkhoi River and at the interfluve of the Sarkhoi and Malaya Kharaganta rivers. Here, we sampled for microphytological analysis not only traditionally macerated dark shales and siltstones, but also clayey varieties of carbonates and cherty dolomites containing lenses and interlayers of black cherts. The collected samples contain organic-walled microfossils in a valley of the right nameless tributary of the Malaya Kharaganta River, feeding into it 5 - 6 km above its mouth. A very poor microfossil set, including only the rare acritarchs *Leiosphaeridia incrassata* (Naum.) and *L.* sp. (20 - 30 μm in diameter) and the sheaths *Eomycetopsis robusta* Schopf (3 - 4 μm wide), were revealed in 5 out of 11 collected samples of black and dark gray siltstones of the Tabinzurta Formation.

The adduced data indicates a presence of microfossil sets, poor both in abundance and in diversity, in the Dibi Formation and the Sarkhoi and Bokson groups (Table 1). The simply-organized acritarchs *Leiosphaeridia* and *Leiominuscula* and the sheaths *Leiotrichoides* and *Eomycetopsis* are typical representatives of these sets. All mentioned taxa belong to the most common components of marine, relatively shallow-water assemblages of the organic-walled microfossils known within a wide time-range from the Late Aphebian to the Vendian and the Early Paleozoic (Pyatiletov, 1978; Vendskaya Sistema, 1985; Ogurtsova, 1985; Microfossils ... , 1989). The above-listed simply-organized acritarchs and sheaths belong to the especially numerous "background" forms in the vast majority of Precambrian microbiotas, being accompanied in the post-Aphebian biotas by more complex microfossils.

The prevalence of only simple acritarchs and sheaths fixed in the number of late Precambrian biotas is usually attributed to the reaction of a biota to a spe-

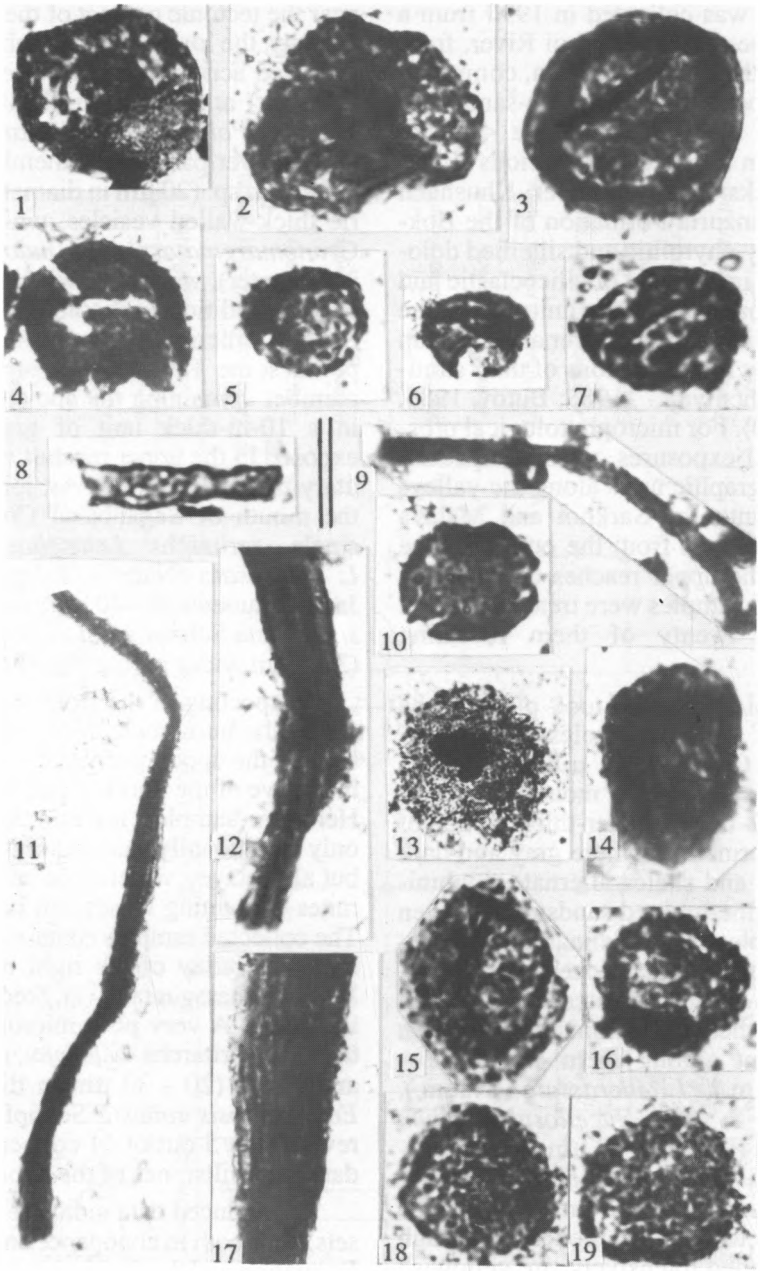

Table 1. *Microfossils of the Dibi Formation; the Sarkhoi and Bokson groups. The Upper Precambrian. Bokson-Sarkhoi depression, southeastern part of the Eastern Sayan.*
All material is kept in the Geological Institute RAN, Moscow.
1 - 3, 5 - 8, 10, 13, 15, 18, 19 – the Dibi Formation; collecting site on nameless creek – right tributary of Urdo-Khundy Creek, joining the latter 3 - 4 km from its mouth.
11, 12, 14, 16, 17 – the Sarkhoi Group, the Khushatai Formation; 11, 12, 16, 17 – collecting site on Sakhor-Zhelga Creek, left tributary of the Sarkhoi River; 14 – collecting site on nameless right tributary of the Sarkhoi River (1.5 - 2 km above the mouth of Saganbaisy Creek). 4, 9 – the Bokson Group, the Tabinzurta Formation; collecting site on the nameless right tributary of the Malaya Kharaganta, joining the latter in 5 - 6 km above its mouth.
Bar equals: 25 μm for 1, 4, 8, 15, 17, 18; 20 μm for 2, 3, 7; 15 μm for 5, 9, 10, 14; 10 μm for 6, 16; 90 μm for 11; 40 μm for 12; and 70 μm for 13, 19.
1 – *Nucellosphaeridium minutum* Tim.; sample N 24902-21. 2, 3, 5, 7, 10, 14 – *Leiosphaeridia crassa* (Naum.); 2 – sample 24902-22; 3 – sample N 24902-23; 5 – sample N 24902-24; 7 – sample N 24902-1; 10 – sample N 24902-26; 14 – sample N 23902-3.
4, 15, 18, 19. *Leiospharidia incrassata* (Naum.); 4 – sample N 25904-1; 15 – sample N 24902-28; 18 – sample N 24902-29; 19 – sample N 24902-30.
6 – *Leiominuscula minuta* Naum.; sample N 24903-2.
8 – *Leiotrichoides typicus* Herm., corroded steath fragment; sample N 24902-25.
9 – *Eomycetopsis robusta* Schopf; sample N 25904-3.
11, 12, 17 – *Polytrichoides lineatus* Hermann; 12, 17 – magnificed images of sample N 21907-3.
13 – *Leiosphaeridia minutissima* (Naum.); sample N 24902-7.
16 – *Granomarginatasphaera judomica* Pjatiletov; small volumetrical thick-walled vesicle; sample N 21908-1.

cific habitat or to the taphonomic peculiarities of microfossils or is interpreted as an artifact appearing as a result of definite factors acting upon the initial samples, causing a selective decomposition of the more vulnerable complex microfossils (e.g., Keller *et al.*, 1984; Veis and Semikhatov, 1989). Taking into account the last data on the environmental settings of organic-walled microfossil assemblages different in their diversity (Vidal and Nystuen, 1990; Veis and Vorob'eva, 1993) and the general lithological features of the sampled Bokson-Sarkhoi depression rocks, it is plausible to relate the poorness of the studied sets of microfossils to the confinement of our samples to those Eastern Sayan paleobasin zones that were characterized by an abrupt impoverishment of the biotas of planktonic and benthic microorganisms (the most distant and extreme proximal shelf zones).

Because of the wide time-range of the majority of the forms present, the studied samples are of little stratigraphic use. The only exceptions are the taxa reported from the upper Sarkhoi Khushatai Formation – the ensheathed multitrichomic filamentous *Polytrichoides lineatus* Hermann and the volumetric thick-walled acritarchs *Granomarginatasphaera judomica* Pjat., known to be restricted to a narrower time-range. The first of these forms is the most distinctive element of the typical Upper Riphean biota of the Urals and Siberia (Muldakayevo, Lakhada, Miroedikha biotas) and rises to the Upper Vendian (Redkino) deposits (Belarus', Moldova), where it appears as a subordinate member of assemblages (Vendian System, 1985). The second form is also known in the Late Riphean-Vendian time-range (the Lakhada Group of the Middle Siberia Uchur-Maya region, the Staraya Rechka, the lower part of the Nemakit-Daldyn formations of the western Anabar region, and others), but it is especially representative and is the most numerous in the Upper Vendian deposits, being one of the important components of the Siberian III-b microfossil complex (Stanevich, 1987, 1988; Pyatiletov, 1988), known in analogues of the upper part of the Moty Group, in the interior regions of the Siberian platform (Khomentovskii *et al.*, 1986).

Thus, our microphytological data strongly suggests the Vendian age of the upper Sarkhoi Khushatai Formation. This conclusion is supported by the above data on isotopic age of the volcanics from the lower part of the Sarkhoi Group (718 ± 70 Ma) and on the composition of stromatolites and archaeocyates (Semikhatov and Serebryakov, 1967) from the above-lying Bokson Group (from the Zabit and Khuzhirtai formations, respectively). Among all earlier known microbiotas of the Vendian age, the Tolparovo microbiota of the Upper Precambrian of the southern Urals, reported from the silicoclastic Tolparovo Formation on the Zilim River (Keller *et al.*, 1984; Mikhailova and Podkovyrov, 1987), is the most similar to the studied upper Sarkhoi microbiota. As for published references on the presence of the Paleozoic organic remains in the deposits of the Bokson-Sarkhoi depression, they are explained, apparently, by contamination of the samples (Khomentovskii, 1985).

Thus, we obtained the first reliable data on the presence of organic-walled microfossils in the Upper Precambrian deposits of the southeastern part of the Eastern Sayan, in the Dibi, upper Sarkhoi (Khushatai), and lower Bokson (Tabinzurta) formations of the Bokson-Sarkhoi depression. The composition of the displayed microfossils in the Dibi and Tabinzurta formations is too primitive for any stratigraphic conclusions, and the forms present in the upper Sarkhoi Khushatai Formation suggest the Vendian age of the enclosing rocks.

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REFERENCES

Butov, Yu.P., Some Problems of the Pre-Mesozoic Stratigraphy of the Sayan-Baikal Mountain Region. Paper 1. A Crisis of Traditional Stratigraphic Scheme, *Byul. MOIP, Otd. Geol.*, 1985, vol. 60, no. 6, pp. 40 - 53.

Butov, Yu.P., Some Problems of the Pre-Mesozoic Stratigraphy of the Sayan-Baikal Mountain Region. Paper 2. New Data on Age of Pre-Mesozoic Formations, *Byul. MOIP, Otd. Geol.*, 1986, vol. 61, no. 2, pp. 91 - 109.

Buyakaite, M.I., Kuz'michev, A.B., and Sokolov, D.D., A 718 Ma Rb-Sr Erochrone of the Sarkhoi Group of the Eastern Sayan, *Dokl. Akad. Nauk SSSR*, 1989, vol. 309, no. 1, pp. 150 - 154.

Dobretsov, N.L., To the Nappe Tectonics of the Eastern Sayan, *Geotektonika*, 1985, no. 1, pp. 39 - 50.

Keller, B.M., Veis, A.F., and Gorozhanin, V.M., The Tolparovo Section of the Upper Precambrian, the South Urals, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1984, no. 9, pp. 119 - 124.

Khomentovskii, V.V., On the Inwash of Small Fossils into Older Formations and Related Stratigraphic Problems, *Geol. Geofiz.*, 1985, no. 1, pp. 6 - 12.

Khomentovskii, V.V., Pak, K.P., Postnikov, A.A., and Skopintsev, V.G., Geological Structure of the Ukha-Gol River Basin (the Eastern Sayan), *Stratigrafiya Pozdnego Dokembriya i Rannego Paleozoya Sibiri*, IGiG SO Akad. Nauk SSSR, Novosibirsk, 1985, pp. 76 - 106.

Khomentovskii, V.V., Shenfil', V.Yu, and Pyatiletov, V.G., To the Problem of Application of the Vendian-Cambrian Microfossil Complexes of Siberia for Correlation and Dating, *Geol. Geofiz.*, 1986, no. 4, pp. 130 - 132.

Kuz'michev, A.B., The Sarkhoi Group of the Stratotype Locality: Stratigraphy, Structural Setting, Age, *Pozdnii Dokembrii i Rannii Paleozoi Sibiri. Voprosy Regional'noi Stratigrafii*, IGiG SO Akad. Nauk SSSR, Novosibirsk, 1990, pp. 104 - 123.

Mikhailova, N.S. and Podkovyrov, V.N., Microphytological Characteristics of the Riphean and Vendian Boundary Formations of the South Urals, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1987, no. 9, pp. 75 - 84.

Mikrofossilii Dokembriya SSSR (Microfossils of the Precambrian in the USSR), Nauka, Leningrad, 1989.

Ogurtsova, R.N., *Rastitel'nye Mikrofossilii Opornogo Razreza Venda-Nizhnego Kembriya Malogo Karatau* (Plant

Microfossils of the Reference Section of the Vendian-Lower Cambrian of the Malyi Karatau Range), ILIM, Frunze, 1985.

Pyatiletov, V.G., Microfossils of the Mana Depression, *Proc. of the IGiG SO Akad. Nauk SSSR*, Novosibirsk, 1978, issue 400, pp. 175 - 185.

Pyatiletov, V.G., Late Precambrian Microfossils of the Uchur-Maya Region, *Pozdnii Dokembrii i Rannii Paleozoi Sibiri. Rifei i Vend*, IGiG SO Akad. Nauk SSSR, Novosibirsk, 1988, pp. 47 - 105.

Resheniya Vsesoyuznogo Stratigraficheskogo Soveshchaniya po Dokembriyu, Paleozoyu i Chetvertichnoi Sisteme Srednei Sibiri (Decisions of the All-Union Conference on the Precambrian, Paleozoic, and Quaternary System of Middle Siberia), Nauka, Novosibirsk, 1983.

Semikhatov, M.A. and Serebryakov, S.N., The Vendian and the Lower Cambrian of the Southeast Part of the Eastern Sayan, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1967, no. 4, pp. 87 - 103.

Stanevich, A.N., Microfossils of the Reference Sections of the Upper Precambrian of the Patom Zone, *Pozdnii Dokembrii i Rannii Paleozoi Sibiri. Sibirskaya Platforma i Ee Yuzhnoe Skladchatoe Obramlenie*, IGiG SO Akad. Nauk SSSR, Novosibirsk, 1987, pp. 30 - 51.

Stanevich, A.N., Stratigraphy and Microfossils of the Late Precambrian of the Patom Zone and North Baikal Region, *Cand. Sci. (Geol.-Min.) Dissertation*, IGiG SO Akad. Nauk SSSR, Novosibirsk, 1988.

Veis, A.F. and Semikhatov, M.A., The Lower Riphean Omakhta Assemblage of Microfossils of Eastern Siberia: Composition and Environmental Setting, *Izv. Akad. Nauk SSSR. Ser. Geol.*, 1989, no. 5, pp. 36 - 55.

Veis, A.F. and Vorob'eva, N.G., Microbiota of the Kerpyl Group of the Siberian Hypostratotype of the Riphean, *Stratigr. Geol. Korrel.*, 1993, vol. 1, no. 1, pp. 41 - 58.

Veis, A.F. and Petrov, P.Yu., Dependence of Systematic Diversity of the Riphean Organic-Walled Microfossil Complexes on Their Environmental Settings (The Bezymyanny Formation of Turukhansk Region as Example), *Ekosistemnye Perestroiki i Evolyutsiya Biosfery*, Nedra, Moscow, 1993, issue 1, in press.

Vendskaya Sistema. Istoriko-Geologicheskoe i Paleontologicheskoe Obosnovanie. Paleontologiya, Tom 1 (Vendian System. Historical-Geological and Paleontological Substantiation. Paleontology, vol. 1), Nauka, Moscow, 1985.

Vidal, G. and Nystuen, J.P., Micropaleontology, Depositional Environment and Biostratigraphy of the Upper Proterozoic Hedmark Group, Southern Norway, *Amer. J. Sci.*, 1990, vol. 290-A, pp. 170 - 211.

Yankauskas, T.V., Mikhailova, N.S., and German, T.N., All-Union Colloquium on the Precambrian Microfossils of the USSR, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1987, no. 9, pp. 137 - 139.

Zhabin, V.V. and Stepanova, M.V., The Lower Boundary of the Yudomian (Vendian) in the Area of the Bokson Mineral Deposit of the Eastern Sayan, *Materialy po Stratigrafii i Paleontologii Sibiri*, IGiG SO Akad. Nauk SSSR, Novosibirsk, 1974, pp. 30 - 33.

Stratigraphy and Geological Correlation, Vol. 1, No. 4, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 4, 1993, pp. 33 - 40.
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Evolution of the biosphere

History of the Evolution of the Stropheodontacea (Brachiopod) in the Devonian Basin of the Western Part of the Altai-Sayan Region

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Abstract – The evolutionary history of the Stropheodontacea (brachiopod) is traced through the Devonian. The changes in time in community composition at the species level are analyzed for the families Stropheodontidae and Douvillinidae, which inhabited the Altai-Sayan basin. Local schemes correlated with the Standard Stratigraphic Scale on conodonts form a stratigraphic base for the analysis. The abrupt boundaries of the reorganizations of brachiopod communities are defined according to the following: (1) complete change in species composition by the beginning of the Emsian; (2) maximum species diversity and abundance in the Emsian; (3) abrupt change of community composition at the Givetian–Frasnian boundary; (4) complete extinction of the Stropheodontacea by the end of the Frasnian, which is correlated with global extinctions in other faunal groups, including brachiopods, at the Frasnian–Famennian boundary.

Levels of extreme changes in the biotic and abiotic environment have been recognized in the Phanerozoic during the last decade. Usually, they are identified well through out the sedimentary basins on different continents, and many of them, globally. The Devonian contains a series of such levels, which are described as global biotic and geological (sedimentological) events (Walliser, 1985; Walliser *et al.*, 1988; House, 1985; Chlupac and Kukal, 1986; and others).

The reorganizations of the ecosystems and their range become most apparent by analyzing paleontological materials through the entire sedimentary basins. We have already attempted to carry out such an analysis, based on brachiopod genus compositions for the single Altai-Salair Devonian basin of the western part of the Altai-Sayan region, against the background of trilobite morphogenesis periodicity, orogenic phases, and eustatic sea-level changes.

The results of such an analysis at the species level of one of the frequently occurring brachiopod groups, such as the Stropheodontacea, are adduced below. The representatives of this group are described monographically in several works (Khalfin, 1931, 1932, 1933a, b, 1937, 1948; Gratsianova, 1960, 1967, 1975; Rzhonsnitskaya, 1968; Alekseeva *et al.*, 1970). The data published in the cited and other works (Yolkin, 1968; Yolkin *et al.*, 1985; Stratotipicheskie ... , 1986, 1987; Gratsianova *et al.*, 1988) form a stratigraphic basis for analysis. The data on conodonts (Rzhonsnitskaya *et al.*, 1975; Gagiev *et al.*, 1986; Yolkin and Izokh, 1988; and others) and our own recently obtained data is used for a more accurate correlation of local stratigraphic schemes with the Standard Stratigraphic Scale (Oliver and Chlupac, 1991).

The superfamily Stropheodontacea is a group of articulate brachiopods that preferred a soft substrate and calm water biotopes. The Devonian reveals such facies in many parts of the Gornyi Altai and in the recently exposed sections near Gur'evsk in the Salair Mountains (Stratotipicheskie ... , 1986, 1987) and the Kuznetsk Basin. The geographical distribution of families and genera composing this superfamily is irregular. Aside from its dependence upon facies, their distribution shows a more or less clear relationship with paleozoogeographical provinces. At the same time, an interprovincial exchange of migrants was active as far as the Stropheodontacea, inhabited mainly in the outer shelf-zone or, at least, at some distance from the sea coast.

In the Devonian of the Altai-Sayan region, the representatives of the Stropheodontidae and Douvillinidae families occur most frequently among the Stropheodontaceae. Therefore, the data on the evolution (appearance, facial control and geographical connections, acme and extinction) of mainly stropheodontids and douvillinids of the Salair Ridge, the Gornyi Altai, and the outskirts of the Kuznetsk Basin will be given below. The

successive evolution of these families at a species level is represented in Fig. 1, and a relative quantitative estimation of their changes in time is shown on a histogram (Fig. 2).

The first representatives of the Stropheodontidae family are known to have appeared in the early Llandoverian, tracing their origin from eoleptostrophians. They are represented by three genera in the Přidolian and Lochkovian (Harper and Boucot, 1978b). Thus, the Devonian, beginning from the Pragian, is a time of acme and extinction of stropheodontids. Moreover, the maximum of their diversity variety (11 genera) corresponds to the Emsian. At the end of the Emsian, 4 of 11 genera died out, and then, at the end of the Eifelian, five of eight genera became extinct (Harper and Boucot, 1978b). The unique phenomenon of a mass (bank) accumulation of thick-shelled specimens of the species *Megastrophia (Megastrophia) uralensis* (Verneuil) in the pure shoal limestone facies in the Urals, the Tien Shan, and the Salair Ridge is attributed to the Emsian (the *serotinus* Zone). Other stropheodontids in reef limestones are represented by very stunted (usually small in size) forms.

Like stropheodontids, Douvillinidae, which appeared in the Wenlockian, is a dominating group (in number of species) of the Stropheodontacea in the Devonian of the Altai-Sayan region. The maximum of their worldwide acme corresponds to the Lochkovian and Pragian (seven genera in each of them); mass extinction took place at the end of the Emsian (10 of 16 genera) and again at the end of the Eifelian (five of eight genera) (Harper and Boucot, 1978a).

Douvillinids had developed two morphological types of shells during adaptation to the substrate: (1) a wide, almost flat or moderately concave-convex, thin-walled shell with thin radial sculpture; (2) a usually large, steeply bent (productidlike) shell with a thickened ventral valve and coarse sculpture of large costellae and shell undas (Gratsianova, 1975, 1987). The species with shells of the first type (the majority of douvillinids) dwelt only on muddy substrates of calm-water parts of the basin. The species of the second type (the genus *Cymostrophia* and the species *Douvillina arcuata* (Hall)) inhabited a soft substrate, being semiburied in it, and, in addition, they adapted themselves to a coarse-grained substrate in an unstable environment near reefs.

A review of the evolution of stropheodontids and douvillinids in the Devonian of the western part of the Altai-Sayan region (the Salair Ridge, the outskirts of the Kuznetsk Basin, and the Gornyi Altai) is given for the regional biostratigraphical superhorizons, which are correlated with the stages of the Standard Stratigraphic Scale (Fig. 1).

The Kaibal Superhorizon (the lower Lochkovian) is subdivided into two horizons. These are the Tom'-Chumysh Horizon in the Salair Ridge, composed of dark bedded limestones with a basal variegated terrigenous member (the Sukhaya Formation), and the Petts Horizon with light massive limestones. Tom'-Chumysh limestones contain an impoverished brachiopod assemblage with *Mesodouvillina (Protocymostrophia) costatula* (Barrande) and *Mesodouvillina (Mesodouvillina) costrialis* (Kozlowski) (Alekseeva *et al.*, 1970). The species *Mesodouvillina (Protocymostrophia) costatula* and *Cymostrophia stephani* (Barrande) are present in Petts limestones as rare specimens.

In the Gornyi Altai, the Kaibal Superhorizon includes the Remnevskii Horizon, consisting of terrigenous rocks barren of brachiopods (below) and dark bedded limestones with an impoverished stunted, small-shelled brachiopod assemblage (above). The widely distributed species *Cymostrophia stephani*, which is scanty here, and the new species *Cymostrophia radiosa* Gratsianova and *Cymostrophia gibbosa* Grats are found (Gratsianova, 1967). The latter species is also known in the Urals, from the Toshemka Horizon of the Pragian, and, in the former Czechoslovakia, it is very close to the Pragian species *Cymostrophia convoluta* (Barr.) (Havlicek, 1967).

Comparison of the Kaibal (the early Lochkovian) Stropheodontacea of the Salair Ridge and Gornyi Altai shows that their evolution in the Salair part of the basin with more open marine environments was gradual: the genus *Mesodouvillina* separated at the beginning of the Lochkovian into two subgenera, i.e., the nominative one and *Protocymostrophia*; at the end of the Lochkovian, the second subgenus was replaced by the genus *Cymostrophia*. Accordingly, a change in the species composition took place; the ancient species *Mesodouvillina (Protocymostrophia) costatula* still existed at the beginning of the early Lochkovian, and the species *Cymostrophia stephani*, which became widely distributed later in the Pragian, appeared at the end of the Lochkovian.

Slightly different environments existed in the Gornyi Altai, where Anuya-Chuya Bay represented a rather narrow, shallow sea. Inner shelf environments, unfavorable for the Stropheodontacea, predetermined an adaptive radiation of the *Cymostrophia* genus. The new species *Cymostrophia radiosa* and *Cymostrophia gibbosa*, mentioned above for the Gornyi Altai, appeared here in the Remnevskii time (the early Lochkovian). Both of them have clearly expressed features of adaptation to a habitat on a soft, friable substrate. Thus, the species *Cymostrophia radiosa*, having a shell with a wide, flat, visceral disk and a geniculately bent train and spines, dwelt on a sediment surface, whereas the shell of the species *Cymostrophia gibbosa*, with a semispherical ventral valve and large wrinkles in intercostal spaces, shows that it was semiburied into the sediments, with its frontal margin elevated above them because of its long train. The coarse sculpture of the valve prevented a full subsidence into the sediment. Each adapting in its own way, both species continued to exist further into the late Lochkovian and Pragian; the species *Cymostrophia gibbosa* also existed in the Urals in the Toshemka time.

The Beltir Superhorizon (the upper Lochkovian and Pragian). It is known that the Devonian was a geocratic

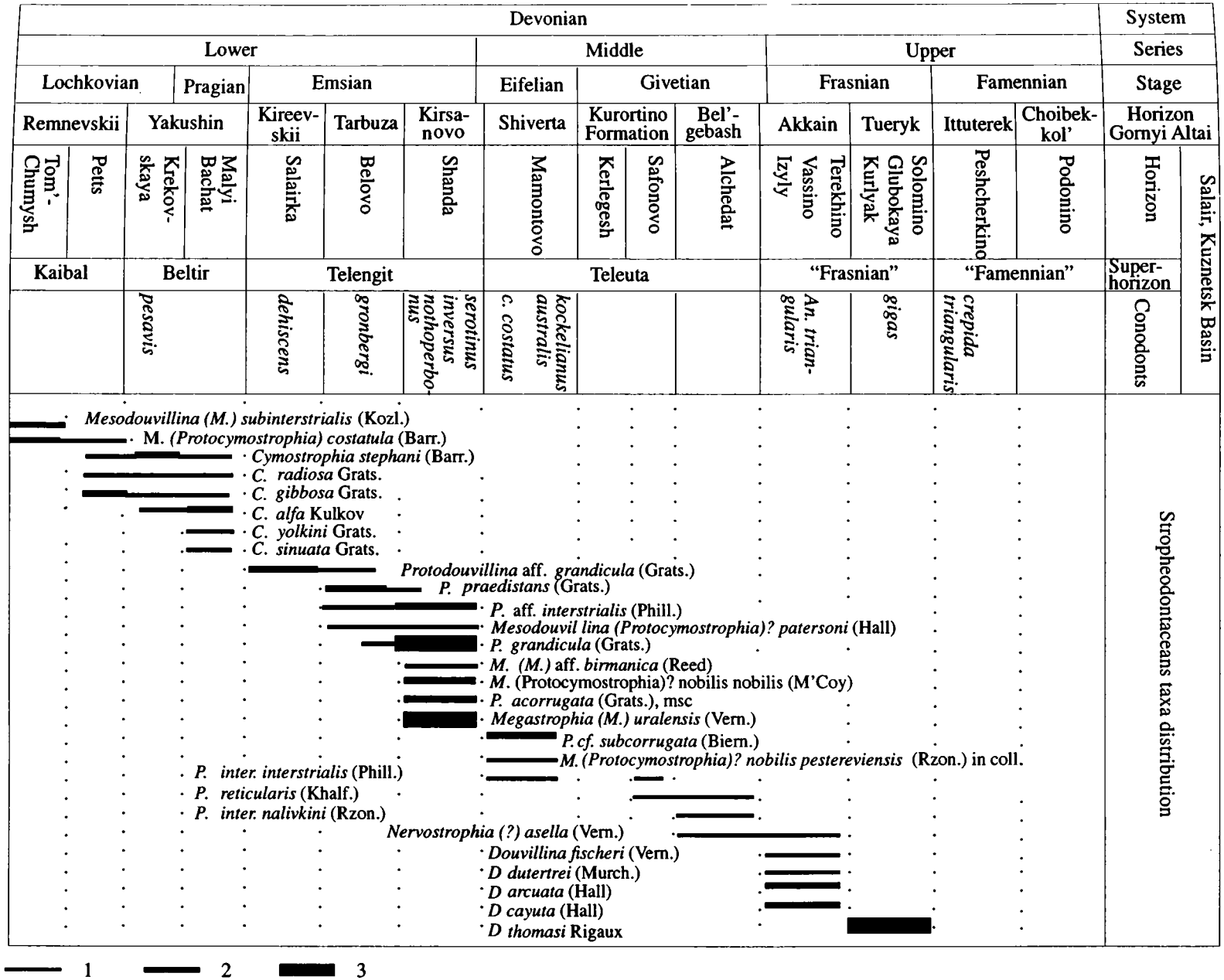

Fig. 1. *Stratigraphic distribution of the Stropheodontacea in the Devonian of the Altai-Sayan region (the thickness of distribution lines reflects relative taxon's abundance): rare (1), common (2), abundant (3).*

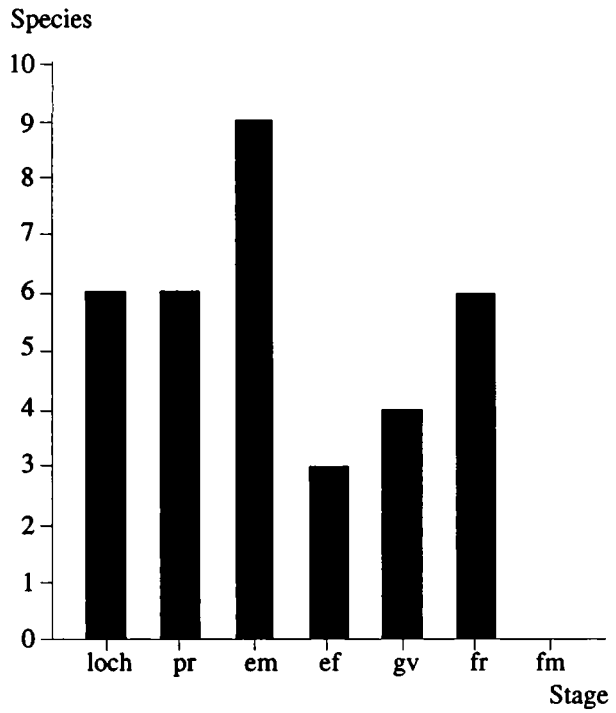

Fig. 2. *Histogram of the Stropheodontacea taxonomic diversity in the stages of the Devonian (compiled using data on the stropheodontid and douvillinid distribution, see Fig. 1). Stages in Latin abbreviation are on horizontal axis, the number of taxa are on vertical axis. Latin abbreviations are names of the stages (loch – Lochkovian, pr – Pragian, em – Emsian, ef – Eifelian, gv – Givetian, fr – Frasnian).*

epoch with sharply expressed climatic belts. The intensive carbonate formation due to the reef buildups of all types took place in tropic basins of this epoch. Leeward slopes of such buildups and distant banks on carbonate platforms provided very favorable conditions for many organisms, including brachiopods. The biotope of the back-reef parts of basins was the best for the Stropheodontacea. An open shelf below a base of obvious instability and a zone of bioherm formation were also colonized by the representatives of this superfamily. The Beltir brachiopod communities in the tropic basins were communities of high diversity, with mainly large-shelled individuals in all facies except those of the reef core. The Stropheodontacea was also taxonomically diverse, though it yielded in abundance to pentamerids, rhynchonellids, and some spiriferids. The Carnian Alps, the Barrandien, the Urals, the Tien Shan, and the Altai-Sayan region represent the most striking examples, in this respect, of the studied age interval.

The Beltir Superhorizon in the Salair is represented by dark bedded limestones of the Krekovskaya Horizon (the upper Lochkovian) in the lower part and by usually light massive limestones of the Malyi Bachat Horizon (the Pragian) in the upper portion. The stropheodontid species *Cymostrophia stephani, Cymostrophia radiosa*, and *Cymostrophia gibbosa* continued to exist here during the Beltir time, becoming extinct by the end of this time. The species *Cymostrophia alfa* Kulkov (Kul'kov, 1960), which is also known from Toshemka limestones of the Urals (Breivel' and Breivel', 1977), appeared in the Krekovskaya time and become extinct by the end of the Malyi Bachat time. The new species *Cymostrophia yolkini* Gratsianova (Alekseeva *et al.*, 1970), with a wrinkled ventral valve, as is typical for leptenids, is found only in the Malyi Bachat limestones. The continuous shell unda, a characteristic feature of the species, favored, apparently, a habitat of the species on a friable substrate in unstable water. This species is noted in the Sauma Horizon (the upper Lochkovian) of the Urals.

The Yakushin Horizon, consisting of terrigenous rocks at the lower part (the upper Lochkovian) and predominantly gray massive limestones in the upper portion (the Pragian), is attributed to the Beltir Superhorizon in the Gornyi Altai. The reef facies of the polyfacies Yakushin limestones is the Solov'ikha limestone (Gratsianova, 1967; Alekseeva *et al.*, 1970), whose age was considered older for a long time (Khalfin, 1948; Kul'kov, 1963; Ivanovskii and Kul'kov, 1974).

The Yakushin stropheodontids are represented by the same species as from the Malyi Bachat, with the addition of a new species. This is *Cymostrophia sinuata* Gratsianova (Gratsianova, 1967), characterized by a small shell with a narrow sinus on the ventral valve and carinate elevation on dorsal valve, which is unusual for cymostrophians. The specimens of the species are collected from reef limestones; the representative features are explained, apparently, by the adaptation of the species to a habitat on a firm substrate in unstable water: two prominences of the sinus sides provided stabilization for the shell.

Thus, the Beltir time (the late Lochkovian–Pragian) in the paleozoogeographical province of the Old World was an epoch during which shelf facies and biotopes were very diverse, which was favorable for brachiopods in general. But the Stropheodontacea was not a polyfacies group, and its attempts to expand into neighboring (respective to traditional) niches initiated with morphological adaptive radiation and the appearance of new specialized species and terminated with their rapid extinction. The species *Cymostrophia yolkini* and *Cymostrophia sinuata* can be regarded as examples of such species with a short biochrone (biozone). These species, as can be expected, are of a very low abundance and can be ascribed as a rarely occurring species; the second species is local one, known only on the northwest of the Gornyi Altai.

The Telengit superhorizon of the Altai-Sayan region, stratigraphically, is approximately equal to the Emsian Stage of the Standard Stratigraphic Scale, including in addition, one short conodont *c. partitus* Zone of the Eifelian Stage. Direct evidence of this, concerning the conodonts, is absent, as the uppermost part of the Telengit Superhorizon consists of nonmarine sediments. However, such a conclusion seems to be the most reliable (Stratotipicheskie ... , 1986), on the basis of the data presently obtained from the sections of this superhorizon in its type locality.

This superhorizon includes the Salairka (the *dehiscens* zone), Belovo (the *gronbergi* Zone), and Shanda (the *nothoperbonus, inversus, and serotinus* zones and, supposedly, the *c. patulus* and *c. partitus* zones) horizons in the Salair Ridge and, corresponding to them, the Kireevskii, Tarbuza (Layer 1 of the Kuvash Beds), and Kirsanovo (Layer 2 of the Kuvash Beds) horizons, and the Malyi Korgon (Gratsianova, 1975), Mukurcherga, and Matveevo horizons (beds) in the Gornyi Altai.

The Salairka Horizon, in its lower part, is represented by predominantly terrigenous rocks; in its middle part, by limestones; and in the upper part, by marls. The rare stropheodontacean specimens *"Gorgostrophia" gorgoides* Gratsianova and *Protodouvillina* aff. *grandicula* (Gratsianova) (Gratsianova, 1975) occur only in dark gray platy limestones of the middle part, with abundant large-shelled pentamerids and spiriferids. The first species has small shells with sharpened auricles and wrinkles on them. The species is rare; the conditions for existence on a muddy substrate are not favorable for it. *"Gorgostrophia" gorgoides* occurs more often in the sandy limestones and sandstones (denser sediment) of the Kireevskii Horizon. It is not known in other regions, appearing here at the beginning of the *dehiscens* time, and becoming extinct by the end of this time.

The species *Protodouvillina* aff. *grandicula* has prove to be more adapted to a muddy substrate in a basin of the Salairka time. Its large, productidlike shell has a long train and wrinkles near the apexes, which provided stability in a semiburied position in a substrate. This species, including both congener forms *P.* aff. *grandicula* and *P. grandicula*, is outwardly homeomorphic to some cymostrophians because of their similar habitat conditions and, like them, it had a rather long biochrone from the *dehiscens* time until the end of the *serotinus* time. The species *Protodouvillina* aff. *grandicula* is rare, having a smaller shell in the Kireevskii Horizon of the Gornyi Altai, where shallow facies prevail.

The Belovo Horizon of the Salair Ridge and the synchronous Tarbuza Horizon of the Gornyi Altai are represented by various facies, consisting of argillites and aleurolites with thin limestone interlayers, thick-bedded organic-detrital limestones, and partly of light massive limestones of algal reef buildups. The species *Protodouvillina praedistans* and *Mesodouvillina (Protocymostrophia)? patersoni* (Hall) are present in all mentioned rock types, but the first of them is more abundant in light limestones, and the second one, in limy aleurolites. *Protodouvillina* aff. *interstrialis* (Phillips) and *Protodouvillina grandicula* appear for the first time as rare forms, whose acme would occur later.

The Shanda Horizon of the Salair Ridge is the most diverse in facies and, accordingly, is characterized by several facies brachiopod communities (Rzhonsnitskaya *et al.*, 1985; Stratotipicheskie ..., 1986, 1987; Yolkin *et al.*, 1988). The Stropheodontacea is abundant only in argillites and interlayering limestones. The following species are abundant: *Protodouvillina* aff. *interstrialis*, which is transitional from the Belovo Horizon; *Protodouvillina grandicula*, connected phylogenetically with the Salairka-Belovo species *P.* aff. *grandicula*; and *Megastrophia uralensis* (Verneuil), appearing for the first time. The latter represents a thin-shelled form of the medium size with well-preserved sculpture and microsculpture. These features are, apparently, typical for this species. Once appearing on a bank with a firm substrate and active hydrodynamics, however, the specimens *Megastrophia uralensis* adapted gradually to such conditions. The bank form mentioned above has a large thick shell with a resorbed surface.

The species *Mesodouvillina (M.)* aff. *birmanica* (Reed) and *Protodouvillina acorrugata* Gratsianova and the subspecies *Mesodouvillina (Protocymostrophia)? nobilis nobilis* (M'Coy) also first occurred on biotopes with soft substrate in the Shanda time. All cited species of the Shanda time became extinct by the end of the Emsian. Only the nominative subspecies *Mesodouvillina (Protocymostrophia)? nobilis* gave rise to the Eifelian subspecies *Mesodouvillina (Protocymostrophia)? nobilis pestereviensis* (Rzonsnitskaya, in coll.).

The Shanda Horizon in the Gornyi Altai is represented by predominantly fine-grained terrigenous and sometimes limy rocks. The Stropheodontacea are less diverse. Some, however, are abundant, including the species *Cymostrophia?* aff. *patersoni*, *Protodouvillina* aff. *interstrialis*, and *P. grandicula*, *Mesodouvillina* aff. *birmanica*. The nonbank form of the species *Megastrophia uralensis* occurs in single specimens.

The Teleuta Superhorizon of the Altai-Sayan region includes the Mamontovo, Kerlegesh, Safonovo, and Alchedat horizons of the Salair Ridge and their analogs in the Gornyi Altai (Fig. 1). The superhorizon corresponds to the stratigraphic interval between two major historical-geological boundaries well-known in Siberia and the Urals, which are located at the base of the conodont *c. costatus* and *Lowermost asymmetricus* zones. The superhorizon in such scope corresponds on the Standard Stratigraphic Scale to the complete Eifelian Stage without its lower *c. partitus* Zone, and Givetian Stage, without its upper *Lowermost asymmetricus* Zone.

The Mamontovo Horizon (the Eifelian), in its type locality in the Salair Ridge, is represented by five facies: (1) sandy-clayey rocks with interlayers and members of black limestones; (2) dark gray bedded limestones; (3) light massive reef limestones; (4) argillites with subordinate aleurolites and sandstones; and (5) inequigranular sandstones with subordinate gravelstones and argillites. These facies are interconnected by vertical and lateral transitions. They are variously distributed in different sections among three independent stratigraphic intervals, which were earlier distinguished as beds (or horizons) with geographical names (from the base to top): the Malaya Salairka Bed consists of the first and second facies with normal stratigraphic relationships between them; the Pesterevo Bed is represented by the third and second facies, which are lateral age analogs to each other; and the Akarachkino Bed is formed by fifth and fourth facies, which are also lateral analogues to each other.

The brachiopod communities show obvious differences in various facies. The first facies contains rare small-shelled brachiopods of low diversity; the second facies includes a rather diverse community with shells of normal sizes; and the third facies is characterized by the diverse and abundant brachiopod assemblage. However, brachiopods of the third facies are evidently stunted, as very small forms are present mainly among them. The latter adapted for existence in a zone of reef buildups, fastening by a podium or leaning against a firm substrate by heavier apexes. Only three stropheodontacean species – *Mesodouvillina (Protocymostrophia)? nobilis pestereviensis* (Rzonsnitskaya, in coll.), *Protodouvillina* cf. *subcorrugata* (Biernat), and *Protodouvillina interstrialis interstrialis* are found in the Pesterevo Limestones. Argillite facies (the Akarachkino Beds) are barren of brachiopods, and sandstone facies, in addition to abundant floral remains, is characterized only by rare finds of brachiopods. Stropheodontacea are absent in these rocks.

The Shiverta Horizon of the Gornyi Altai is an age analogue of the Mamontovo Horizon. It is represented at the base by terrigenous rocks almost barren of fossils in the lower part and by carbonate rocks with abundant faunal remains, including brachiopods, in the upper portion. Stropheodontacea are not found here.

The Kerlegesh and Safonovo horizons of the Salair Ridge compose the lower Givetian. In type locality in the Salair Ridge, the first of them consists of dark bedded limestones, and the second one, mainly of reef limestones. A considerable amount of tuffaceous material is present in the sections of the northwest Salair Ridge. Stropheodontacea are recorded only in Safonovo limestones.

The Safonovo limestone brachiopod assemblage varies both in composition and in abundance. The species diversity is not high in general, and the Stropheodontacea are rare. The species *Protodouvillina interstrialis interstrialis* (Phill.) is transitional from the Eifelian, and *Protodouvillina reticularis* (Khalfin) (Khalfin, 1937) passes into the upper Givetian; both are found in clayey limestones. The first of them is geographically very widespread; the second one is confined to Siberia only.

The early Givetian in the Salair Ridge was not favorable for the Stropheodontacea. The finds of organic remains in the Gornyi Altai are hardly known at all for this interval, inasmuch as it is represented by effusive rocks (the Kurotino Formation). Pillow lavas prevail among volcanics, thus creating extremely unfavorable conditions for fauna in general in the Altai-Salair basin.

The Alchedat Horizon (the late Givetian), in the outskirts of the Kuznetsk Basin, and its age analog in the Gornyi Altai (Bel'gebash Horizon) consist entirely of extremely shallow-water formations of the late Givetian transgression. The facies of this time are variable and mosaic, both in the stratigraphic section and laterally. The species composition of brachiopods changed completely. The appearing communities of so-called "Chiel'" fauna were characterized by low diversity, with a dominance of large-shelled, coarse-sculptural forms inhabiting shallow water. The Stropheodontacea are rare here, being poor in composition and in abundance. Only three forms are known: the Siberian species *Protodouvillina reticularis*, transitional from the Safonovo Horizon; younger subspecies of the early Givetian species *Protodouvillina interstrialis nalivkini* (Rzonsnitskaya); and the Urals species *Nervostrophia* (?) *asella* (Verneuil), appearing here and transitioning into the lower Frasnian.

The Frasnian seas of the outskirts of the Kuznetsk Basin and the Gornyi Altai, as well as Givetian seas, were shallow and unstable. The facial picture was also variable, and brachiopod communities were also of low diversity; however, there is some difference in the history of brachiopod evolution during the two ages being compared. After the occurrence of intensive volcanic activity in the early Givetian, a highly impoverished brachiopod assemblage made a final attempt to survive. For example, the athyridids burst of genus- and species-formation (Khalfin, 1933a; Gratsianova and Dagis, 1983) is illustrative in this respect. Some evolutionary activity is also observed for the stropheodontacean genus *Protodouvillina*, which gave rise to the new genus *Douvillina*, with a number of species: *D. dutertei* (Murchison), *D. arcuata* (Hall), and *D. fischeri* (Verneuil), in the early Frasnian, and *D. thomasi* Rigaux, in the late Frasnian. All mentioned species occur in one or another section of the Vassino Horizon of the Kuznetsk Basin and in the Akkain and Tueryk horizons of the Gornyi Altai (Fig. 1). The history of evolution of the Stropheodontacea superfamily terminated at the end of the late Frasnian; no species had passed the Frasnian-Famennian boundary anywhere.

Thus, the following abrupt boundaries of reorganization in communities (Figs. 1, 2) are outlined in the evolutionary history of the Stropheodontacea in the Devonian of the Altai-Sayan region: (1) a complete change of community composition at the boundary of the Beltir and Telengit superhorizons (at the very beginning of the Emsian); (2) maximum diversity and abundance of the species in the Telengit Superhorizon (the Emsian) and in the Shanda Horizon (the late Emsian or Dalean) in particular; (3) an abrupt change in community composition at the Givetian-Frasnian boundary.

The established significant impoverishment of the Givetian Stropheodontacea assemblages correlates with the occurrence of intensive subaqual volcanism in the basin. All stropheodontids became extinct by the end of the Frasnian, which corresponded to the global extinction in other faunal groups, including brachiopods, at the Frasnian–Famennian boundary.

REFERENCES

Alekseeva, R.E., Gratsianova, R.T., Yolkin, E.A., and Kul'kov, N.P., *Stratigrafiya i Brakhiopody Nizhnego Devona Severo-Vostochnogo Salaira* (Stratigraphy and Brachiopods of the Lower Devonian of Northwestern Salair), Nauka, Moscow, 1970.

Breivel', I.A. and Breivel', M.G., Brachiopods, *Biostratigrafiya i Fauna Rannego Devona Vostochnogo Sklona Urala*, Nedra, Moscow, 1977, pp. 52 - 105, 152 - 157, 163 - 173.

Chlupac, I. and Kukal, Z., Reflection of Possible Global Devonian Events in the Barrandian Area, C.S.S.R., *Global Bio-Events. Lecture Notes in Earth Sciences*, no. 8, Berlin, 1986, pp. 169 - 179.

Gagiev, M.Kh., Rodygin, S.A., and Timofeeva, O.B., Conodont Complexes and Zonal Subdivision of the Devonian of the Salair Ridge, *Novye Dannye po Stratigrafii Dokembriya i Paleozoya Altae-Sayanskoi Skladchatoi Oblasti*, Zapsibgeologiya, Novokuznetsk, 1986, pp. 49 - 51.

Gratsianova, R.T., Brachiopods of the Eifelian of the Korgon Range in the Gornyi Altai, *Voprosy Stratigrafii i Paleontologii Zapadnoi Sibiri*, Izd. Akad. Nauk SSSR, Novosibirsk, 1960, pp. 139 - 152.

Gratsianova, R.T., *Brakhiopody i Stratigrafiya Nizhnego Devona Gornogo Altaya* (Brachiopods and Stratigraphy of the Lower Devonian of the Gornyi Altai), Nauka, Moscow, 1967.

Gratsianova, R.T., *Brakhiopody Rannego i Srednego Devona Altae-Sayanskoi Oblasti. Strofomenidiny* (Brachiopods of the Early and Middle Devonian of Altai-Sayan Region. Strophominidina), Nauka, Moscow, 1975.

Gratsianova, R.T., Levels of Brachiopod Habitat with Respect to the Basin Floor, *Sreda i Zhizn' v Geologicheskom Proshlom. Paleolandshafty Morei i Kontinentov*, Nauka, Novosibirsk, 1987, pp. 41 - 58.

Gratsianova, R.T. and Dagis, A.S., Morphology and Problems of the Systematics of Some Devonian Athyridids of Western Siberia, *Morfologiya i Sistematika Bespozvonochnykh Fanerozoya*, Nauka, Moscow, 1983, pp. 15 - 27.

Gratsianova, R.T., Talent, J.A., and Yolkin, E.A., Devonian Brachiopod Zonal Schemes, Provinciality and Crustal Blocks in the Asia-Australia Hemisphere, *Devonian of the World. Proc. of the Second Intern. Symposium on the Devonian System*, Canad. Soc. Petrol. Geologist, Calgary, Canada, vol. III, 1988, pp. 295 - 301.

Harper, C.W. and Boucot, A.J., Stropheodontacea. Part II, *Paleontographica, Part A*, 1978a, vol. 161, nos. 4 - 6, pp. 119 - 175.

Harper, C.W. and Boucot, A.J., Stropheodontacea. Part III, *Paleontographica, Part A*, 1978b, vol. 162, nos. 1 - 2, pp. 1 - 80.

Havlicek, V., Brachiopoda of the Suborder Strophominidina in Czechoslovakia, *Rozpr. Ústředn. Ústavu Geol.*, 1967, vol. 33, pp. 1 - 235.

House, M.R., Correlation of Mid-Paleozoic Ammonoid Evolutionary Events with Global Sedimentary Perturbations, *Nature*, 1985, vol. 313, pp. 119 - 122.

Ivanovskii, A.B. and Kul'kov, N.P., *Rugozy, Brakhiopody i Stratigrafiya Silura Altae-Sayanskoi Gornoi Oblasti* (Rugoses, Brachiopod and Stratigraphy of the Silurian of the Altai-Sayan Region), Nauka, Moscow, 1974.

Khalfin, L.L., Upper Devonian Brachiopods of the Cherepanov Ford on the Yaya River, *Materialy k Stratigrafii Severnoi Okrainy Kuznetskogo Ugol'nogo Basseina, Part 2*, 1931, pp. 1 - 32.

Khalfin, L.L., Lower Frasnian Brachiopods of the Outskirts of the Kuznetsk Coal Basin and Gorlovka Coal-Bearing Region, *Izv. Zap.-Sib. Geol.-Razv. Tresta*, 1932, vol. XII, no. 2, pp. 1 - 48.

Khalfin, L.L., The Genus *Anathyris* Reetz in the Devonian of Siberia, *Sbornik po Geologii Sibiri, Posvyashchennyi M.A. Usovu*, TII, Tomsk, 1933a, pp. 108 - 132.

Khalfin, L.L., The Upper Devonian near Zharkovskoe Village on the Yaya River, *Materialy k Stratigrafii Severnoi Okrainy Kuzbassa*, GNTGI, Novosibirsk, 1933b.

Khalfin, L.L., Middle Devonian Brachiopod of Lebedyanskoe Village, Anzhero-Sudzhensk Region, *Izv. Tomsk. Industr. Inst.*, 1937, vol. 57, no. 1, pp. 85 - 148.

Khalfin, L.L., Fauna and Stratigraphy of the Devonian of the Gornyi Altai, *Izv. Tomsk. Politekh. Inst.*, 1948, vol. 65, no. 1.

Kul'kov, N.P., *O Faune Brakhiopod Pesterevskikh Izvestnyakov i Ikh Fatsial'nykh Analogov* (About Brachiopod Fauna of Pesterevo Limestones and Their Facial Analogs), *Proc. Inst. Geol. and Geophys.*, SO Akad. Nauk SSSR, 1960, no. 1, pp. 153 - 207.

Kul'kov, N.P., *Brakhiopody Solov'ikhinskikh Sloev Nizhnego Devona Gornogo Altaya* (Brachiopods of the Solov'ikha Beds of the Lower Devonian of the Gornyi Altai), Izd. Akad. Nauk SSSR, Moscow, 1963.

Oliver, W.A., Jr. and Chlupac, I., Defining the Devonian: 1979 - 89, *Lethaia*, 1991, vol. 24, no. 1, pp. 119 - 122.

Rzhonsnitskaya, M.A., *Biostratigrafiya Devona Okrain Kuznetskogo Basseina. T. 1. Stratigrafiya* (Devonian Biostratigraphy of the Kuznetsk Basin Outskirts. Vol. 1. Stratigraphy), Nedra, Leningrad, 1968.

Rzhonsnitskaya, M.A., Gagiev, M.Kh., Kulikova, V.F., Spasskii, Ya.N., and Timofeeva, O.B., The Eifelian-Givetian Boundary in the Salair Ridge, *Srednii Devon SSSR, Ego Granitsy i Yarusnoe Raschlenenie*, Nauka, Moscow, 1985, pp. 132 - 139.

Stratotipicheskie Razrezy Nizhnego i Srednego Devona Salaira. Telengitskii Nadgorizont: Terrigenno-Karbonatnye Fatsii (Stratotypical Sections of the Lower and Middle Devonian of the Salair Ridge. The Telengit Superhorizon: Terrigenous-Carbonate Facies), Yolkin, E.A. and Kanygin, A.V., Eds., IGiG, Novosibirsk, 1986.

Stratotipicheskie Razrezy Nizhnego i Srednego Devona Salaira. Telengitskii Nadgorizont: Karbonatnye Fatsii (Stratotypical Sections of the Lower and Middle Devonian of the Salair Ridge. The Telengit Superhorizon: Carbonate Facies), Yolkin, E.A. and Kanygin, A.V., Eds., IGiG, Novosibirsk, 1987.

Walliser, O.H., Natural Boundaries and Commission Boundaries in the Devonian, *Cour. Forsch.-Inst. Senckenberg*, 1985, vol. 75, pp. 401 - 407.

Walliser, O.H., Lottmann, J., and Schindler, E., Global Events in the Devonian of the Kellerwald and Harz Mountains, *Cour. Forsch.-Inst. Senckenberg*, 1988, vol. 102, pp. 190 - 193.

Yolkin, E.A., *Trilobity (Dekhenellidy) i Stratigrafiya Nizhnego i Srednego Devona Yuga Zapadnoi Sibiri* (Trilobites (Dechenellids)) and Stratigraphy of the Lower and Middle Devonian of the South of Western Siberia), Nauka, Moscow, 1968.

Yolkin, E.A., Bakharev, N.K., Gratsianova, R.T., and Zheltonogova, V.A., To the More Precise Definition of the Stratigraphic Scheme of the Lower Devonian on the Northeast Slope of the Salair Ridge, *Biostratigrafiya i Biogeografiya Paleozoya Sibiri*, IGiG, Novosibirsk, 1985, pp. 76 - 98.

Yolkin, E.A. and Izokh, N.G., Zonal Conodont Species in the Type Localities of the Telengit Superhorizon (Devonian, Salair), *Fauna i Stratigrafiya Paleozoya Srednei Sibiri i Urala*, Nauka, Novosibirsk, 1988, pp. 3 - 17.

Yolkin, E.A., Gratsianova, R.T., Bakharev, N.K., *et al.*, Facies and Faunal Association of the Telengitian (Emsian) in Its Type Locality, *Devonian of the World, V. III: Paleontology, Paleoecology and Biostratigraphy*, McMillan N.J., Embry A.F., and Glass D.J., Eds., Canad. Soc. Petrol. Geologist, Calgary, 1988, pp. 193 - 207.

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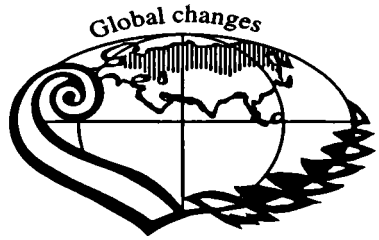

Evolution of the biosphere

Stages of Development of the Triassic Palynofloras and Their Relation to the Changes in Paleogeographical Environments

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Abstract – Generalization of the palynological data on the Triassic deposits of the Moscow, Pechora, and Caspian syneclises, the western Caucasus, Syria, and some Western European regions presented the possibility of tracing three stages of palynoflora development, coinciding in general with the stages determined on plant macrofossils. The first stage includes the Early Triassic and Anisian palynocomplexes, with some elements of the Late Permian palynoflora. The second stage embraces the Ladinian-Carnian palynoflora, characterized by essential changes in its composition due to the new forms, presages of the Mesophytic, appearing in the Ladinian, and Paleophytic forms becoming extinct. The third stage corresponds to the Norian-Rhaetian palynoflora, when the taxa of previous stages disappear, and the real Mesophytic species, which will define the Jurassic palynoflora, assume prominence.

The Triassic palynoflora, as the entire Triassic, being at the boundary of two eras, occupies a dual position. If the Early Triassic is like a natural continuation of the Late Permian, the Late Triassic is connected closely with the Early Jurassic. The Early Triassic and Anisian palynocomplexes, each having individual features, preserved in many respects the Permian features, but in the Ladinian, they were essentially changed in connection with the appearance of new forms, presages of the Mesophytic flora, which were supplanted by the real Mesozoic specimens in the Late Triassic and became dominating in the Jurassic. Thus, changes of palynoflora reflect a long transition from the Paleophytic to the Mesophytic, which covers a period from the Late Permian to the middle of the Triassic; i.e., the real Paleophytic-Mesophytic boundary falls approximately on the Ladinian, when the important rearrangement in palynoflora composition took place.

Palynological research of the Triassic deposits of the Moscow, Pechora, and Caspian syneclises, eastern Mediterranean region (the western Caucasus and Syria), and published data made it possible to outline the main stages of palynoflora evolution (Fig. 1). The stages, based directly on certain palynological material, coincide in general features with the stages defined on plant macrofossils (Dobruskina, 1982; Meien, 1987). This paper deals with the palynoflora evolutionary stages and other paleobiogeographical features of the Triassic.

The first stage includes the Early Triassic and Anisian palynoflora, which contain many forms common with the Permian complexes, preserving the Paleophytic features. An arid climate existed in Eurasia in the Late Permian to Middle Triassic, according to V.M. Sinitsyn (1980). Plant fossils and palynomorphs are indicators of climate features, along with lithological types of sedimentary rocks and animal world associations. The major Permian regression, reaching its maximum by the end of this period, had intensified the climate dryness and thereby stimulated the development of xerophile flora, very poor in species, which has been reflected in the character of the following Early Triassic palynoflora.

Pollen (with ribbed and unribbed bodies), resembling that of Peltaspermaceae, Voltziaceae, and other coniferous plants, plays a noticeable role among the Triassic miospores from the red continental deposits of the first half of the Lower Triassic in the Moscow and Pechora syneclises. The majority of such pollen in the Permian, as well as in the Triassic, belongs to protosaccate (Scheuring, 1974) or to quazisaccate (Meien, 1987) forms. Pollen, represented by both ribbed (*Striatopodocarpites*, *Taeniaesporites* (*Lunatisporites*, *and Striatohaploxypinites*) and unribbed (*Triadispora*, *Klausipollenites*, *Voltziapites*, and *Vesicaspora*) forms in association with the spores of pteridophytic and lycopsidic plants, predominates in the Lower Triassic Vokhma Horizon palynocomplexes of the Moscow syneclise (Goman'kov *et al.*, 1986). The majority of the

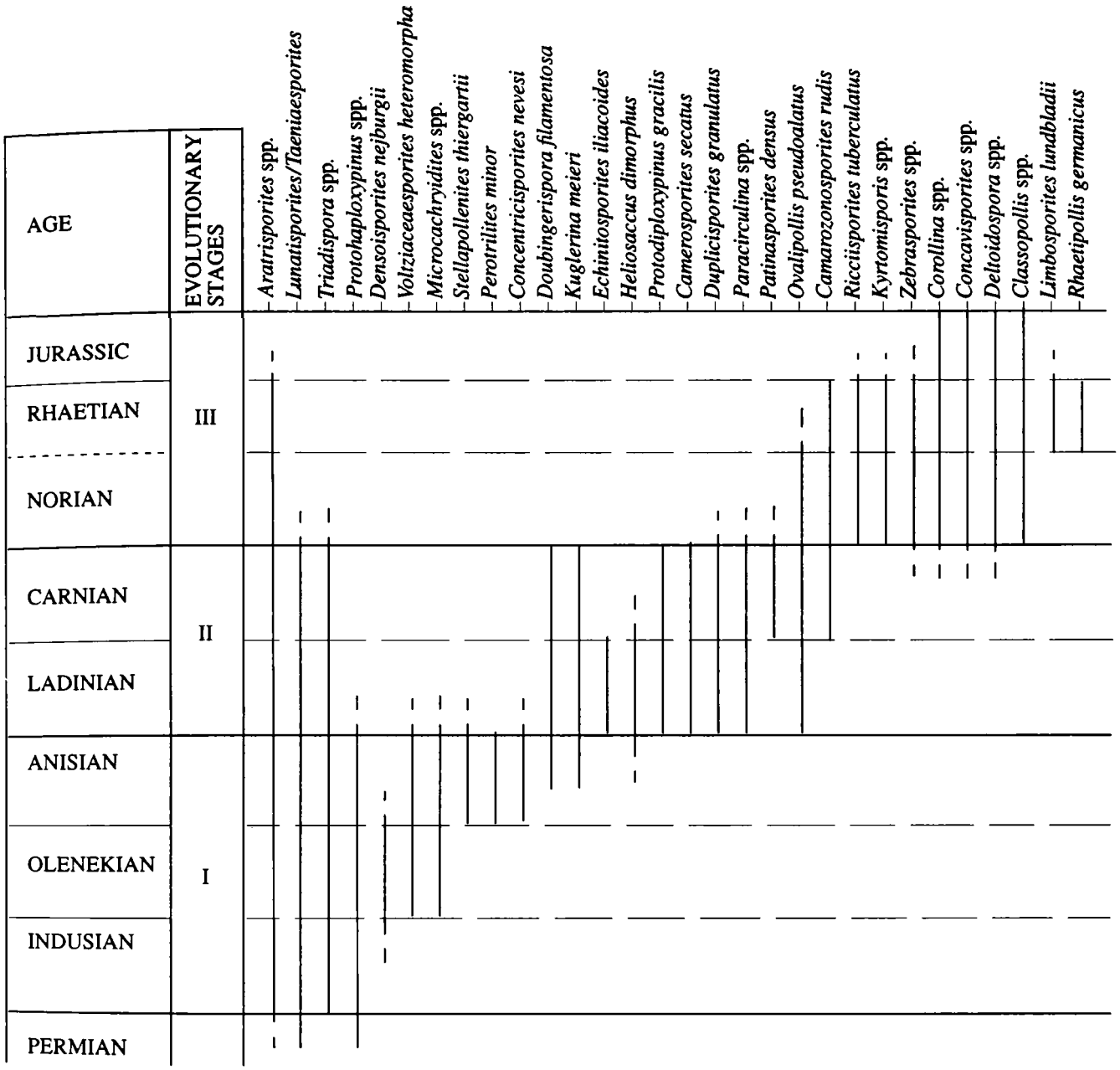

Fig. 1. *Distribution of main miospore taxa reflecting the stages of Triassic palynoflora evolution.*

pollen was typical for the Tatarian Stage. Pteridophytic and lycopsidic spores are numerous in the Lower Triassic sections of the Pechora syneclise, where pollen plays a minor role and is represented by the *Protohaploxypinus*, *Taeniaesporites*, *Striatopodocarpites*, and *Striterspovites* genera (Yaroshenko *et al.*, 1991). The complexes of the first half of the Lower Triassic exhibited some similarity to those from the western sections studied in the Griesbach deposits of eastern Greenland (Balme, 1979) and Arctic Canada (Fisher, 1979) and from the lower Buntsandstein of western Poland (Orłowska-Zwolinska, 1984), though the low variety of taxonomic composition is characteristic of the latter.

In the second half of the Early Triassic, the cavate spores of lycopsids of the *Densoisporites* genus, belonging to Pleuromeiaceae, which became almost globally distributed in both marine and continental deposits, came to the foreground, along with *Taeniaesporites* and other miospores. Olenekian deposits of the Moscow syneclise (Goman'kov *et al.*, 1986; Kyuntt-sel', 1966), Mt. Bol'shoe Bogdo (Sinegub, 1972), the western Caucasus (Yaroshenko, 1978), and Libya (Adloff *et al.*, 1986); middle Buntsandstein rocks of Germany (Reitz, 1985; Schulz, 1964), western Poland (Orłowska-Zwolinska, 1977), and the Spathian deposits of Italy (Visscher, 1974); and the Smithian rocks of

the Barents Sea (Hochuli *et al.*, 1989) contain the most representative occurrences of these spores.

The maximum spread of Pleuromeiaceae coincided with the Olenekian marine transgression embracing the vast areas that are both adjacent to the Boreal basin and situated north of the Tethys. The main Pleuromeiaceae finds have been confined to the Tethys and Boreal basin coasts, and the location of these remains, in the center of the German basin, corresponds to a coastal line of the advancing sea. The latter indicates that in the Early Triassic, in connection with an intense aridization, no plants could survive on the marine coasts except Pleuromeiaceae (Dobruskina, 1982). Mass finds of *Densoisporites* spores are usual in the layers with aquatic invertebrates and amphibians. The aridization, increasing in the Middle Triassic, promoted a wide distribution of the gymnosperm pollen *Lunatisporites* (*Taeniaesporites*), *Triadispora*, *Angustisulcites*, *Illinites*, *Microcachryidites*, and *Voltziaceaesporites*, among which the forms inherited from the Permian played a definite role. Such finds, in association with typical Anisian miospores, are connected predominantly with carbonate deposits of Western Europe (Adloff and Doubinger, 1977; Doubinger and Adloff, 1981; Klaus, 1964; Reitz, 1985; Taugourdeau-Lantz, 1974), Libya (Adloff *et al.*, 1986), and Syria (Yaroshenko and Bash Imam, 1989).

The second stage, including the Ladinian and Carnian palynoflora, is marked by an essential change of its composition, connected with an appearance in the Ladinian of the new pollen group *Circumpolles*, represented by its most ancient genera (*Camerosporites*, *Duplicisporites*, and *Paracirculina*) in the association of typical miospores. The appearance of the *Circumpolles* group, whose acme was connected with the Carnian Stage, was a biotic event caused in the Ladinian, obviously, by the evolution of the flora from Voltziaceae to Cheirolepidiaceae. This concept is based on the data of Scheuring (1976), who studied in detail the morphology of the pollen of *Circumpolles* (*Camerosporites* and *Duplicisporites*), *Triadispora*, and *Doubingerispora* and defined the phylogenetic relations of the pollen genus *Triadispora*, resembling Voltziaceae (Grauvogel-Stamm, 1978), and pollen of the early *Circumpolles*, the ancestor of the *Classopollis* genus, belonging to the Cheirolepidiaceae (Vakhrameev, 1970, 1980). Specimens of protomonosaccate pollen of the *Doubingerispora* and, apparently, the *Kuglerina* genera (possessing the morphological features of both *Triadispora* and *Circumpolles*, which existed simultaneously) are considered to be transition forms between the pollen of the above-mentioned genera. Both pollen groups belong to a category of xerophile forms, indicators of an arid climate (Eshet and Cousminer, 1986; Visscher and Van der Zwan, 1981). The pollen of *Camerosporites secatus*, *Duplicisporites*, *Paracircula*, *Ovalipollis*, *Triadispora*, *Patinasporites densus*, *Doubingerispora filamentosa*, *Kuglerina meieri*, and other taxa, found together, made it possible to identify the *Camerosporites secatus* phase for the Ladinian and Carnian stages (Besems, 1983; Van der Eem, 1983; Visscher and Krystyn, 1978; Visscher and Van der Zwan, 1981).

The development of the *C. secatus* phase in the Alpine sections is regarded as a considerable event in the evolution of Middle and Upper Triassic palynoflora. Its representatives were widespread, not only in the Alpine and German sections of Europe, but far beyond its limits as well, from Arctic Canada to northwestern Australia.

The palynocomplexes belonging to the *C. secatus* phase in sections of the Mediterranean region, situated in the equatorial arid zone, showed the possibility of revealing paleogeographical and paleoclimatic features (Visscher and Van der Zwan, 1981). Proportions of hydrophile and xerophile miospores, defined for coal-bearing and carbonate deposits, suggest the climate changing from temperate and humid in the east, where the hydrophile categories prevailed, to arid in the west, with a dominance of xerophile groups. The latter is caused, apparently, by a westward tendency of the humidity decrease, from the western Caucasus (Yaroshenko, 1978) to Spain (Besems, 1981a, b; 1983), situated in the extreme west of the Tethys in the most arid area of this region, which was reflected both in the xerophile palynoflora and in evaporites.

A general aridization, continuing in the Ladinian and, particularly, in the Carnian in Europe and the Middle East, was manifested in the xerophile character of the typical Mediterranean palynofloras. These included the palynocomplexes with a predomination of *Circumpolles*, *Triadispora*, and *Lunatisporites* and the participation of other miospores, which were recorded in southern France (Taugourdeau-Lantz, 1984), Italy (Van der Eem, 1983), Spain (Besems, 1981a, b; 1983), Libya (Adloff *et al.*, 1986), Israel (Eshet and Cousminer, 1986), and Syria (Yaroshenko and Bash Imam, 1989). In contrast to them, the complexes with a dominance of hydrophile categories of various pteridophytic spores, along with numerous lycopsids (*Aratrisporites*), are also known in this time. These are the palynocomplexes from coal-bearing deposits of Lunz in Austria (Bharadwaj and Singh, 1964), Germany (Reitz, 1985), Poland (Orłowska-Zwolinska, 1977, 1979), and the northern Caspian region of Russia (Shelekhova and Yasamanov, 1988; Yaroshenko, 1969). Associations from the Barents Sea sections (Hochuli *et al.*, 1989), unlike those of the southern sections, contain an insignificant number of *Circumpolles* pollen and have been referred to hydrophile palynofloras, which was determined by their situation in the Boreal Region.

Thus, in the time of general xerophytization of the floras of the stage under consideration, the hydrophile complexes, reflecting the climatic changes, were rather noticeable, being closely related to the Middle Triassic expansion of the transgression, which began in the Olenekian. This resulted in a humid climate, the formation of gray-colored sediments, the beginning of coal formation, and other phenomena. The transgression was accompanied by the wide distribution of numerous

spores and acritarchs, which were typical of the Ladinian palynocomplexes of Germany and Poland and were replaced by abundant *Circumpolles* pollen in the Carnian.

The third stage of palynoflora evolution fell on the post-Carnian Time, i.e., on the Norian-Rhaetian. Here, I adduce the generalized characteristics of these subdivisions, as the study and discussion of the problem of the Norian and Rhaetian Stages is not yet concluded. The replacement of the ancient elements of the *Circumpolles* group by its young representatives, connected with the appearance the *Classopollis* pollen and with more frequent finds of *Corollina* pollen, is the most important event of the stage under consideration. In addition, the Norian-Rhaetian palynofloras underwent a subsequent change, manifested in a noticeable decline or disappearance of *Camerosporites secatus*, along with other earlier representatives of the *Circumpolles* group, the protosaccate pollen group of *Doubingerispora*, *Kuglerina*, *Triadispora*, and of ribbed forms.

The appearance of the *Classopollis* pollen is remarkable in its being a real Mesozoic flora element, which, together with pteridophytic spores, will determine the character of Jurassic and younger miospore complexes. Besides the *Classopollis* and *Corollina* genera, the diagnostic miospores of *Riccisporites tuberculatus*, *Ovalipollis pseudoalatus*, and *Rhaetipollis germanicus* and the *Zebrasporites*, *Kyrtomisporis*, *Concavisporis*, and *Deltoidospora genera* are inherent in the Norian-Rhaetian. Their main spectrum has a wide distribution in palynofloras of this time, from Greenland (Pedersen and Lund, 1980), Arctic Canada (Fisher, 1979), England (Fisher, 1972), and Poland (Orłowska-Zwolinska, 1983), to Spain (Baudelot and Taugoudeau-Lantz, 1986) and Portugal (Adloff *et al.*, 1974). The *Classopollis* pollen, being a product of the thermophilic and xerophile coniferous plants of the Cheirolepidaceae family, noticeably reacts against the transition from southern arid areas to the more humid northern areas (Vakhrameev, 1970). This is shown by the Greenland (Pedersen and Lund, 1980) and Barents Sea (Hochuli *et al.*, 1989) palynocomplexes, where the *Classopollis* pollen is rare. In contrast to them, this pollen can be traced in large amounts and often is dominant in the central regions of West Europe and the Mediterranean. Palynocomplexes of Germany (Brenner, 1986), Austria (Karle, 1984), Poland (Orłowska-Zwalinska, 1979), Spain (Baudelot and Taugourdeau-Lantz, 1986), Lybia (Adloff *et al.*, 1986), and Syria (Yaroshenko and Bash Imam, 1989) confirm this conclusion. The presence of various pteridophytic spores indicates, apparently, signs of some climatic humidization, which began at the end of the Triassic and continued in the Jurassic.

Finally, one can draw the following conclusions. The Triassic palynoflora is of a dual character. It contains Late Permian representatives in the Early Triassic, whereas in the Late Triassic, its composition is closely connected with Jurassic elements. The outlined stages of the palynoflora evolution correspond in general features to the stages of the flora development defined according to plant macrofossils. The first stage palynoflora (the Early Triassic-Anisian) is characterized by the complexes bearing Paleophytic features. The second stage (the Ladinian-Carnian) is marked by the appearance of new pollen of the *Circumpolles* group, which was a forerunner of the real Mesophytic. Thus, the Ladinian palynoflora is regarded as a crucial border in flora evolution, when the Paleophytic was replaced by the Mesophytic. The third stage (the Norian-Rhaetian) of palynoflora evolution was marked by the appearance of real representatives of the Mesozoic flora, i.e., of the *Classopollis* genus, replacing the early elements of the *Circumpolles* group.

REFERENCES

Adloff, M.C., Doubinger, J., and Palain, C., Contribution à la Palynologie du Trias et du Lias Inférieur du Portugal "Gres de Silves" du Nord du Tage, *Grum. Serv. Géol. Portugal*, 1974, vol. 58, pp. 91 - 144.

Adloff, M.C. and Doubinger, J., Observations Palynologique dans le Muschekalk Inférieur du Sontage de Mersch, Luxemburg, *Bull. Serv. Géol. Luxemburg*, 1977, no. 8, pp. 1 - 15.

Adloff, M.C., Doubinger, J., Massa, D., and Vachard, D., Trias de Tripolitaine (Libye) Nouvelles Données Biostratigraphiques et Palynologique, *Rev. Inst. Fr. Pétrole*, 1986, vol. 41, no. 1, pp. 27 - 72.

Balme, B.E., Palynology of Permian. Triassic Boundary Beds at Kap Stosch, East Greenland, *Medd. Grønland*, 1979, vol. 200, no. 6, pp. 1 - 37.

Baudelot, S. and Taugourdeau-Lantz, J., Dicouverte d'une Microflora dans les Pyrenees Catalanes Attribuable au Norien-Rhetien, *Rev. Paléobiologie*, 1986, vol. 5, no. 1, pp. 5 - 9.

Besems, R.E., Aspects of Middle and Late Triassic Palynology. 1. Palynostratigraphical Data from the Chiclana de Segura Formation of the Linares-Alcarez Region (Southeastern Spain) and Correlation with Palynological Assemblages from the Iberian Peninsula, *Rev. Palaeobot. Palynol.*, 1981a, vol. 32, pp. 257 - 273.

Besems, R.E., Aspects of Middle and Late Triassic Palynology. 2. Preliminary Palynological Data from the Hornos-Siles Formation of the Prebetic Zone, NE Province of Jaen (Southeastern Spain), *Rev. Palaeobot. Palynol.*, 1981b, vol. 32, no. 4, pp. 389 - 400.

Besems, R.E., Aspects of Middle and Late Triassic Palynology. 3. Palynology of the Hornos-Siles Formation (Prebetic Zone, Province of Jaen, Southern Spain), with Additional Information on the Macro- and Microfaunas, *Schriftenreihe der Erdwisseschaftlichen Kommissionen*, 1983, vol. 5, pp. 37 - 56.

Bharadwaj, D.C. and Singh, H.P., An Upper Triassic Miospora Assemblage from the Coals of Lunz, Austria, *Paleobotanist*, 1964, vol. 12, no. 2, pp. 28 - 44.

Brenner, W., Bemerkungen zur Palynostratigraphie der Rhät-Lias Grenze in SW-Deutschland, *Neues Jahbuch Paläont. Abh. Stuttgart*, 1986, vol. 173, no. 2, pp. 131 - 166.

Dobruskina, I.A., *Triasovaya Flora Evrazii* (Triassic Flora of Eurasia), Nauka, Moscow, 1982.

Doubinger, J. and Adloff, M.C., Précisions Palynologiques sur l'Age Anisien Moyen du Gisement a Plantes du Trias des

Lavaredes (Se de Lodéve S de France), *Sci. Géol. Bull.*, 1981, vol. 34, no. 4, pp. 239 - 243.

Eshet, Y. and Cousminer, H.L., Palynozonation and Correlation of the Permian-Triassic Succession in the Negev, Israel, *Micropaleontol.*, 1986, vol. 32, no. 3, pp. 193 - 214.

Fisher, M.J., The Triassic Palynofloral Succession in England, *Geoscience and Man.*, London, 1972, vol. 4, no. 1, pp. 101 - 109.

Fisher, M.J., The Triassic Palynofloral Succession in the Canadian Arctic Archipelago, *AASP Contrib.*, 1979, Ser. 58, pp. 83 - 100.

Goman'kov, A.V., Kyunttsel', M.K., and Meien, S.V., *Novye Dannye o Rastitel'nykh Ostatkakh iz Verkhnei Permi i Nizhnego Triasa Russkoi Platformy* (New Data on Plant Remains from the Upper Permian and Lower Triassic of Russian Platform), Deposited in and available from VINITI, Moscow, 1986, no. 7994-B.

Grauvogel-Stamm, L., *La Flore du Crés a Voltzia (Buntsandstein Supérieur) des Vosges de Nord France, Memoire*, 1978, no. 50.

Hochuli, P.A., Colin, J.P., and Vigran, J.Os., Triassic Biostratigraphy of the Barents Sea Area, *Correlation in Hydrocarbon Exploration. Norwegian Petrol. Society*, Graham and Trotman, London, 1989, pp. 131 - 153.

Karle, U., Palynostratigraphische Untersuchung eines Rhät-Lias Profils am Fonsjoch, Achesee (Nördliche Kalkalpen, Österreich), *Mitt. Österr. Geol. Ges.*, 1984, vol. 77, pp. 331 - 353.

Klaus, W., *Zur Sporenstratigraphischen Einstufung von Gipsführenden Schichten in Bohrungen Heft 4*: Wien, Hamburg, 1964, pp. 3 - 16.

Kyunttsel', M.K., New Data on the Find of *Pleuromeia Rossica Neub.* in the Lower Triassic Deposits of the Northeastern Regions of the Russian Craton, *Znachenie Palinologicheskogo Analiza dlya Stratigrafii i Paleofloristiki*, Nauka, Moscow, 1966, pp. 95 - 96.

Meien, S.V., *Osnovy Paleobotaniki* (Fundamentals of Paleobotany), Nedra, Moscow, 1987.

Orłowska-Zwolinska, T., Palynological Correlation of the Bunte and Muschekalk in Selected Profiles from Western Poland, *Acta Geol.*, 1977, vol. 27, no. 4, pp. 417 - 430.

Orłowska-Zwolinska, T., Miospory, T. *III. Atlas Skamieniałosci Przewodnich i Charakterystycznych. 2a. Mezozoic. Trias*, Malinowskiej, L., Ed., Budowa Geologiczna Polski., Warszawa, 1979, pp. 159 - 201.

Orłowska-Zwolinska, T., *Palynostratigrafia Epikontinentalnych Osadów Wyzszego Triasu w Polsce*, Prace Instytutu Geologicznego, Warszawa, 1983, vol. 104, pp. 5 - 89.

Orłowska-Zwolinska, T., Palynostratigraphy of the Buntsandstein in Sections of Western Poland, *Acta Paleontol. Pol.*, 1984, vol. 29, no. 3/4, pp. 161 - 194.

Pedersen, K.R. and Lund, J.J., Palynology of the Plant-Bearing Rhaetian to Hettangian Kap Stewar Formation, Scoresby Sund, East Greenland, *Rev. Palaeobot. Palynol.*, 1980, vol. 31, pp. 1 - 69.

Reitz, E., Palynologie der Trias in Nordhessen und Südniedersachsen, *Geol. Abh. Hessen*, 1985, vol. 89.

Scheuring, B.W., "Protosaccate" Strukturen ein Weitverbreitete Pollenmerkmal zur Früchen und Mittleren Gymnospermenzeit, *Geol. Paläontol. Mitt. Insbruck*, 1974, vol. 4, no. 2, pp. 1 - 30.

Scheuring, B.W., Proximal Exine Filaments, a Widespread Feature among *Triassic Protosaccites* and *Circumpolles* to Secure the Dispersal of Entire Tetrads, *Pollen et Spores*, 1976, vol. 18, no. 4, pp. 611 - 639.

Schulz, E., Sporen und Pollen aus dem Mitteleren Buntsandstein des Germanischen Bleckens, *Monatsber. Dt. Akad. Wiss.*, 1964, vol. 6, no. 8, pp. 597 - 606.

Shelekhova, M.N. and Yasamanov, N.A., Climatic Conditions on the Territory of the North Caspian Basin during the Second Half of the Triassic According to Palynological Data, *Byul. MOIP, Otd. Geol.*, 1988, vol. 63, no. 2, pp. 90 - 95.

Sinegub, V.P., Leaf Flora, Spore, and Pollen, *Stratotipicheskii Razrez Baskunchakskoi Serii Nizhnego Triasa Gory Bol'shoe Bogdo*, Saratov Univ., Saratov, 1972, pp. 3 - 164.

Sinitsyn, V.M., *Vvedenie v Paleoklimatologiyu* (Introduction to Paleoclimatology), Nedra, Leningrad, 1980.

Taugourdeau-Lantz, J., Premier Étude des Spores du Trias Moyen de Gabian (Bordure Sud de la Montagne Noire, France), *Rev. Palaeobot. Palynol.*, 1974, vol. 17, nos. 1 - 2, pp. 149 - 159.

Taugourdeau-Lantz, J., Les Associations Palynologiques du Trias Languedocien dans Lœur Carde Européen: Influence du Milieu (Sol de Végétation-Milieu Sédimentaire), *Geol. France*, 1984, nos. 1 - 2, pp. 23 - 28.

Vakhrameev, V.A., Distribution Regularities and Paleoecology of the Mesozoic Coniferous Cheirolepidiaceae, *Paleontolog. Zh.*, 1970, no. 1, pp. 19 - 34.

Vakhrameev, V.A., *Classopollis* Pollen as an Indicator of the Jurassic and Cretaceous Climate, *Sov. Geol.*, 1980, no. 2, pp. 48 - 56.

Van der Eem, J.G.L.A., Aspects of Middle and Late Triassic Palynology. 6. Palynological Investigations in the Ladinian and Lower Carnian of the Western Dolomites, Italy, *Rev. Palaeobot. Palynol.*, 1983, vol. 39, pp. 189 - 300.

Visscher, H., The Impact of Palynology on Permian and Triassic Stratigraphy in Western Europe, *Rev. Palaeobot. Palynol.*, 1974, vol. 17, no. 1/2, pp. 5 - 19.

Visscher, H. and Krystyn, L., Aspects of Late Triassic Palynology. 4. A Palinological Assemblage from Ammonoid-Controlled Late Carnian (Tuvalian) Sediments of Sicily, *Rev. Palaeobot. Palynol.*, 1978, vol. 26, nos. 1 - 4, pp. 93 - 112.

Visscher, H. and Van der Zwan, C.J., Palynology of the Circum-Mediterranean Triassic: Phytogeographical and Palaeoclimatological Implications, *Geol. Rundschau*, 1981, vol. 70, pp. 625 - 634.

Yaroshenko, O.P., Miospores in Continental Deposits of the Triassic in the North Caspian Basin and Their Significance for Stratigraphy, *Sov. Geol.*, 1969, no. 5, pp. 64 - 73.

Yaroshenko, O.P., *Kompleksy Miospor i Stratigrafiya Triasa Zapadnogo Kavkaza* (Miospore Complexes and Stratigraphy of the Triassic of the Western Caucasus), Nauka, Moscow, 1978.

Yaroshenko, O.P. and Bash Imam, I., Palinoflora and Stratigraphy of the Middle and Upper Triassic Deposits of Syria, *Paleofloristika i Stratigrafiya Fanerozoya*, GIN Akad. Nauk SSSR, Moscow, 1989, pp. 32 - 65.

Yaroshenko, O.P., Golubeva, L.P., and Kalantar, I.Z., *Miospory i Stratigrafiya Nizhnego Triasa Pechorskoi Sineklizy* (Miospores and the Lower Triassic Stratigraphy of the Pechora Syneclise), Nauka, Moscow, 1991.

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Zonal Subdivision of the Middle Jurassic-Lower Cretaceous in Canada and Eastern Europe according to Foraminifer Evidence

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Abstract – Detailed microbiostratigraphic research is based on species biozone analysis and the establishment of their stratigraphic range. Comparing the data on biozones, the distribution of species common for different regions is elucidated, which enables using them for the study of basins, paleoecology, and paleobiogeography.

INTRODUCTION

This paper deals with the study of the foraminiferal stratigraphic distribution in the Jurassic and Early Cretaceous sediments of the Canadian Atlantic shelf and the East European craton. A great deal of new summarizing data on the Jurassic and Lower Cretaceous foraminifers and stratigraphy of different regions in the Northern Hemisphere has been published lately. The zonal subdivision of the Jurassic and Lower Cretaceous sediments of the Canadian Atlantic coast, Eastern and Western Europe, the Crimea, and the Caucasus was substantiated according to different foraminifer groups (Grigelis and Kuznetsova, 1987; Kuznetsova and Gorbachik, 1985; Stratigrafiya i Korrelyatsiya ... , 1985; Ascoli, 1988, 1990; Williams *et al.*, 1990). However, the stratigraphic range of the Jurassic and Early Cretaceous common foraminifer species was studied using only the data from the Canadian Nova Scotian shelf and Western Europe (Ascoli, 1976). Most of the Middle-Upper Jurassic age-diagnostic species and all of those from the Lower Cretaceous were established to have been distributed both on the Nova Scotian shelf and in Germany. It was also recorded that in many cases, the upper limit of the stratigraphic distribution (point of "extinction") of age-diagnostic species, from the Middle Jurassic, Bathonian, to the Latest Cretaceous, Maestrichtian, was the same on the Nova Scotian shelf and in Germany, as well as in other West European regions.

At present, this data can be supplemented and specified, based on our collections and other published records. We thank the leaders of the Atlantic Geoscience Center and the Lithuanian Geological Institute for presenting the opportunity of joint research, and our colleagues for their critical review of the manuscript and helpful remarks.

METHODS

The studied collections were obtained from deep boreholes on the Canadian Atlantic coast, outcrops and holes from the East European craton (Baltic, Pechora, Moscow, and Saratov regions), and, partly, from the Poland and German basins. In the European regions, the stratigraphic frame is based on ammonites; in the Canadian Atlantic shelf, on calpionellids (in the late Tithonian-middle Valanginian interval); and in both regions, on planktonic foraminifers.

The zonal subdivision of the Jurassic and Lower Cretaceous sediments according to foraminifers was made in the studied regions using somewhat different criteria. On the Canadian Atlantic shelf, the foraminifer and ostracode zones are characterized by the last occurrences of the species, because the material was derived from sludge. Such species compose a zonal diagnostic assemblage. One or two zonal species among them determine the name of a zone. The established zones have a fixed upper boundary and are classified as informal assemblage zones (Ascoli, 1988).

The zonal subdivision of the Jurassic sediments in the East European craton and the Lower Cretaceous sediments in the Crimea is based on ammonites. The study of foraminifers from the same sections made it possible to establish both the first and last occurrence of species. Therefore, the lower and upper boundaries of zones were substantiated. They are classified as assemblage zones, or concurrent-range zones (Prakticheskoe Rukovodstvo ... , 1991; Grigelis, 1985).

Species stratigraphic range. The Jurassic and Lower Cretaceous foraminifer species distribution on the Canadian Atlantic shelf and in the East European craton has not yet been analyzed. Meanwhile, during that time, these regions belonged to a single paleobiogeographic province and were not separated by the North Atlantic. The resemblance of taxonomic faunal composition (foraminifer genera and families) was the result of sim-

ilar paleobiogeographic conditions, as well as of uniform terrigenous-carbonate marine sedimentation.

MIDDLE JURASSIC

Figure 1 shows the stratigraphic range of the Middle Jurassic species on the Canadian Atlantic shelf and in the East European craton. Fifty-three common species were recorded in the Bajocian-Callovian, with a maximum number of 11 species in the lower Bathonian.

Garantella rudia Kaptarenko is a common species in the Bajocian. The Bajocian was penetrated or reached only in several boreholes on the Canadian Atlantic shelf. Nevertheless, the find of *Garantella rudia* permits distinguishing there a new corresponding zone (Erider M-75 borehole, Grand Banks). In the East European craton, the *Garantella rudia* zone has been identified in the Dnieper-Donets basin, where it corresponds to the upper Bajocian *Garantiana garantiana* ammonite zone (Pyatkova and Permyakova, 1978). On the Canadian Atlantic shelf, the *Garantella rudia* zone is dated as the Bajocian. The associated foraminifer assemblage contains rare *Garantella ampasindavayensis* Espitalie et Sigal, *G. semiornata* (Schwager), and *Epistomina praecursor* Ohm (collection of P. Ascoli).

The Bathonian yields seven common species. The planktonic foraminifer *Conoglobigerina bathoniana* (Pazdro) was identified there. In the East European craton, it was recorded in the Bathonian of the Polish Lowland (Pazdro, 1969). The topotypes of *Conoglobigerina bathoniana* from Ogrodzec, Poland (collection of A. Grigelis) are identical with the finds from the Canadian shelf. This species marks the common zonal level.

Among the Bathonian benthic foraminifers, common for both regions, the following forms were found with some difference in the stratigraphic range: *Epistomina bireticulata* Pazdro, *Garantella ornata* (Hofker), *Reinholdella crebra* Pazdro, *Lenticulina daphne* Bielecka et Styk, and *Trocholina conica* (Schlumberger). In the East European craton, these species are known from the Bathonian of the Polish Lowland (western part of the craton), whereas part of them appeared in the late Bajocian (Bielecka and Styk, 1981). In addition to the Bajocian-Bathonian of Poland, *Reinholdella media* (Kaptarenko) was found in the upper Bajocian of the Ukraine (*Strenoceras subfurcatum* and *Garantiana garantiana* zones).

The diagnostic species composition of the *Epistomina bireticulata-Garantella ornata* Zone of the Bathonian on the Canadian Atlantic shelf corresponds to that of the Polish Lowland Bathonian, where we distinguish for the first time the *Epistomina bireticulata-Reinholdella crebra* Zone. In the eastern part of the East European craton, the Bathonian sediments (except for the upper Bathonian) are distributed in the Ryazan'-Saratov and Dnieper-Donetsk basins, but the specific composition is different (Prakticheskoe Rukovodstvo ... , 1991; Pyatkova and Permyakova, 1978).

Seven common species occur in the *Callovian*. The Canadian Atlantic shelf sediments contain the planktonic species *Globuligerina calloviensis* K. Kuznetsova, which permits us to distinguish for the first time the corresponding zone (Acadia K-62 borehole, Scotian shelf). The species is unknown in the Callovian of the East European craton. The zone was identified in the lower Callovian of the Crimea, as well as in the Eastern Mediterranean (Syria). However, finds of this important species are extremely rare.

The stratigraphic range of *Epistomina porcellanea* Brückmann, one of the common benthic species, is limited by the Callovian. Other species, such as *Epistomina coronata* Terquem and *Epistomina regularis* Terquem, appear on the Canadian Atlantic shelf and, earlier in the East European craton, in the late Bajocian or in the Bathonian. On the other hand, *Epistomina nuda* Terquem and *Ophthalmidium carinatum* Kübler et Zwingli are known in the Polish Lowland sediments from the late Bajocian. None of the *Epistomina* species were recorded above the Callovian upper boundary except for *Epistomina mosquensis* Uhlig, which appeared in the Bathonian and was distributed in the Callovian, whereas in the East European craton, it still occurred in the early and middle Oxfordian (in the Polish Lowland) and on the Canadian shelf throughout the Oxfordian and Kimmeridgian.

Epistominids began to dominate on the Canadian shelf in the Callovian. They were the most important foraminifer faunal component throughout the Late Jurassic and the Earliest Cretaceous. Numerous lenticulinids (together with polymorphinids in the early Callovian) were also substantial in the East European craton in the Callovian.

The Callovian zones of the Canadian Atlantic shelf and East European craton are based on different diagnostic assemblages. Among the calcareous benthic diagnostic species of the *Epistomina coronata-Epistomina regularis-Epistomina omninoreticulata* Zone, known from the Callovian of the Canadian shelf, the lower Callovian sediments of the Polish Lowland yield *Ophthalmidium carinatum* Kübler et Zwingli, *Epistomina nuda* Terquem, and *E. coronata* Terquem (Bielecka and Styk, 1981; Budowa Geologiczna ... , 1980). The middle and upper Callovian sediments in the eastern regions of the East European craton contain only *Epistomina porcellanea* Brückmann (Prakticheskoe Rukovodstvo ... , 1991). Within the Bathonian-Callovian agglutinated benthos *Trocholina conica* zone of the Canadian Atlantic shelf, the following species of the *Trocholina* genus occurred in the East European craton: *T. nana* Kaptarenko, in the Bathonian and lower Callovian, *T. conica* (Schlumberger), in the Bathonian, and *T. klaipedica* Grigelis, in the middle and upper Callovian (Grigelis, 1985).

UPPER JURASSIC

Figure 2 illustrates the stratigraphic range of the Upper Jurassic foraminifer species, where 33 common

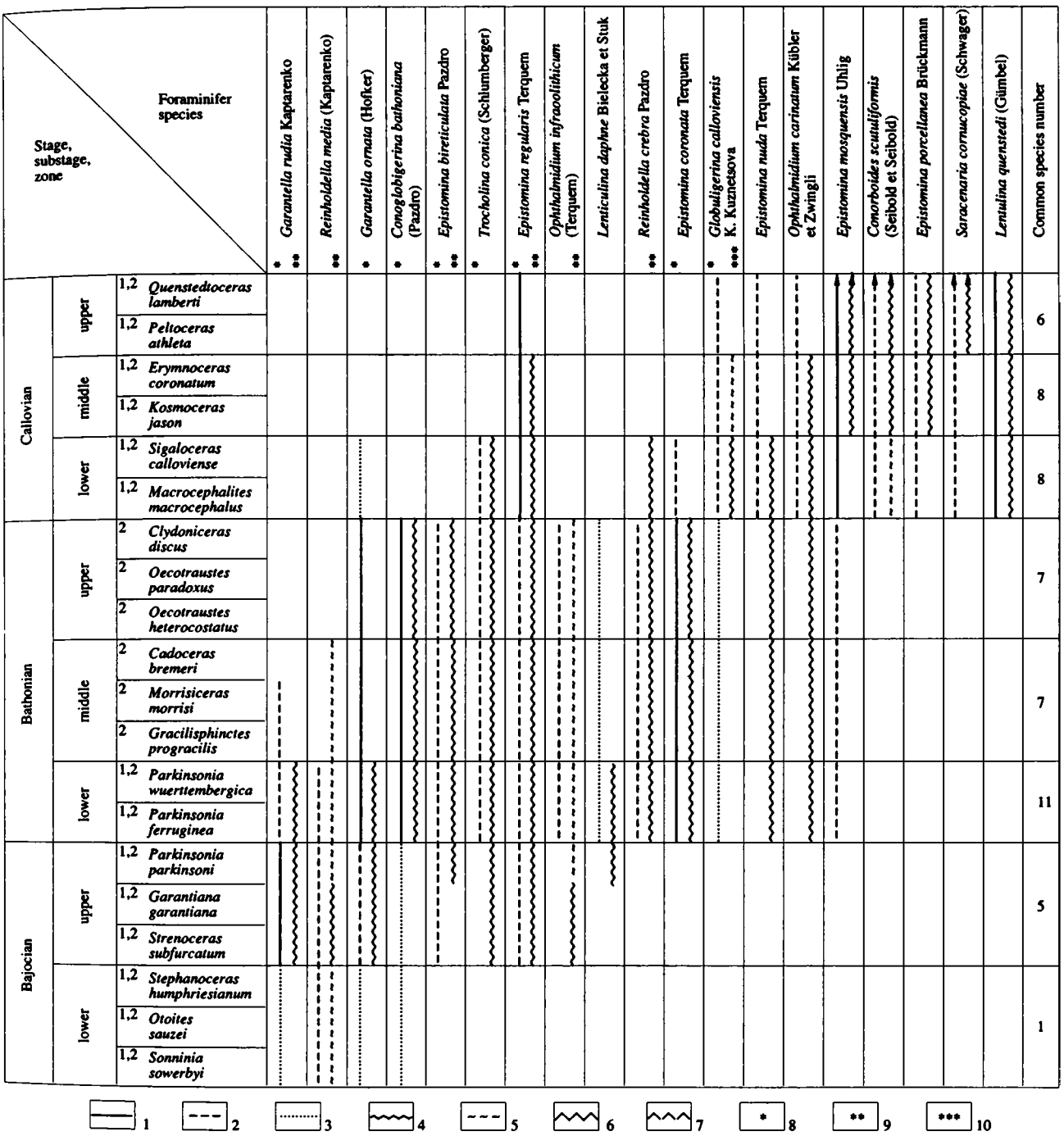

Fig. 1. *Analysis of the stratigraphic distribution of the Middle Jurassic foraminifers on the Canadian Atlantic shelf and in the East European craton.*

Zonal subdivision by ammonites (figures in the column): 1 – East European craton, 2 – Polish Lowland.

Distribution of selected foraminifer species: 1 - 3 – Canadian Atlantic shelf (1 – >5 localities; 2 – <5 localities; 3 – possible occurrence); 4, 5 – East European craton (4 – established occurrence; 5 – possible occurrence); 6, 7 – Southern margin of the East European craton (6 – established occurrence; 7 – possible occurrence). Zonal index species: 8 – Canadian Atlantic shelf; 9 – East European craton; 10 – southern margin of the East European craton.

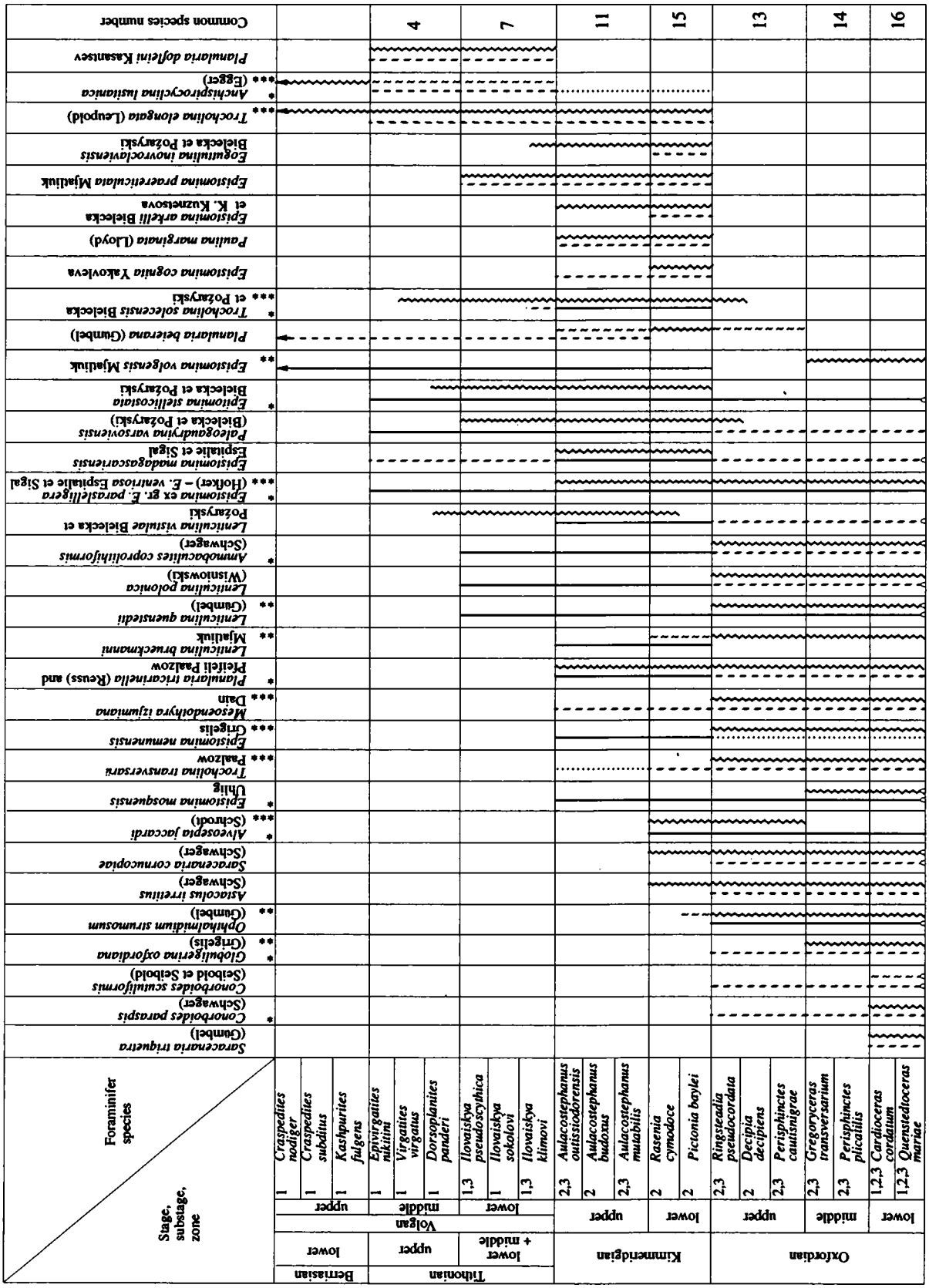

Fig. 2. *Analysis of the stratigraphic distribution of the Upper Jurassic foraminifers on the Canadian Atlantic shelf and in the East European craton.*

Zonal subdivision by ammonites (figures in the column): 1 – East European craton; 2 – southern margin of the East European craton; 3 – Polish Lowland. For symbols see Fig. 1.

forms were recorded within the Oxfordian-Tithonian. Their maximum diversity (about 13 - 16 species) occurred in the Oxfordian and lower Kimmeridgian; the minimum (four species), in the upper Tithonian.

The Oxfordian yields nine common species. The presence of the planktonic form *Globuligerina oxfordiana* (Grigelis) in both regions permits us to distinguish the corresponding zone of the same age (Grigelis and Kuznetsova, 1987, Ascoli, 1988).

The common benthic species include *Saracenaria triquetra* (Gümbel) and *Conorboides paraspis* (Schwager), which are limited by the Oxfordian in both regions. *Conorboides scutuliformis* (Seibold et Seibold) and *Saracenaria cornucopiae* (Schwager) appeared in the Callovian and were characteristic of the Oxfordian. The latter also occurred in the lower Kimmeridgian of the East European craton (Lithuania). The middle Oxfordian zonal species of the East European craton, *Ophthalmidium strumosum* (Gümbel), on the Canadian Atlantic shelf, was found in the Callovian as well. The Oxfordian sediments of both regions contain *Astacolus irretitus* (Schwager) and *Trocholina transversarii* Paalzow, which are also known from the lower Kimmeridgian.

On the Canadian Atlantic shelf, the Oxfordian was defined as the *Epistomina soldanii-Conorboides paraspis* Zone and the *Alveosepta jaccardi* Zone (upper Oxfordian), according to calcareous and agglutinated benthos, respectively. In the East European craton, the Oxfordian zonal subdivision is more detailed and is based on the study of reference sections dated by ammonites (Srednii i Verkhnii Oksford ... , 1989). The *Ophthalmidium sagittum-Epistomina volgensis* Zone of the lower and the beginning of the middle Oxfordian has no common species in the zonal diagnostic assemblages, but it yields *Epistomina nemunensis* Grigelis and *Trocholina transversarii* Paalzow, which appeared on the Canadian Atlantic shelf in the Oxfordian. The East European zonal lower Oxfordian forms *Lenticulina brueckmanni* Mjatliuk and *Epistomina volgensis* Mjatliuk appeared on the Canadian shelf only in the Kimmeridgian. The zonal form *Ophthalmidium strumosum* (Gümbel) is a common species in the diagnostic assemblage of the East European middle Oxfordian *Ophthalmidium strumosum-Lenticulina brestica* Zone. *Epistomina volgensis* Mjatliuk, *E. nemunensis* Grigelis, and *Trocholina transversarii* Paalzow are also present. The East European upper Oxfordian. *Lenticulina russiensis-Epistomina uhligi* Zone yields a common species *Ophthalmidium strumosum* in the zonal diagnostic assemblage. *Astacolus irretitus* (Schwager), *Lenticulina quenstedti* (Gümbel), and *Epistomina nemunensis* Grigelis were also recorded in both regions.

On the Canadian Atlantic shelf and at the southern margin of the East European craton in the Crimea-Caucasus region (Prakticheskoe Rukovodstvo ... , 1991; Ascoli, 1988), the agglutinated benthos *Alveosepta jaccardi* Zone is distinguished as an acme zone at the same level of the upper Oxdordian.

The Kimmeridgian sediments contain 14 common species. Planktonic foraminifers are absent.

The Kimmeridgian age is reliably recorded by the occurrence of calcareous benthic *Epistomina cognita* Yakovleva, *E. arkelli* Bielecka et K. Kuznetsova, and *Paulina marginata* (Lloyd), which are not known in the younger sediments. *Epistomina nemunensis* Grigelis, *Planularia tricarinella* (Reuss), *P. feifeli* Paalzow, and *Lenticulina brueckmanni* Mjatliuk passed from the Oxfordian and became extinct in the Kimmeridgian. *Epistomina mosquensis* Uhlig and *E. volgensis* Mjatliuk were distributed in the Kimmeridgian of the Canadian shelf later than in the East European craton, where they are unknown above the lower-middle Oxfordian. *Epistomina praereticulata* Mjatliuk and *Eoguttulina inovroclaviensis* Bielecka et Pozaryski appeared in both regions in the Kimmeridgian. *Lenticulina vistulae* Bielecka et Pozaryski, characteristic of the Kimmeridgian of Poland, appeared on the Canadian Atlantic shelf in the Oxfordian, whereas *Trocholina solecensis* Bielecka et Pozaryski was found in older sediments in Poland (the upper Oxfordian).

The agglutinated benthic *Alveosepta jaccardi* (Schrodt) still occurred in the Kimmeridgian of both regions. Its upper limit on the Canadian Atlantic shelf and in Poland corresponds to the lower Kimmeridgian, and in the Crimea, to the upper Kimmeridgian. *Mesoendothyra izjumiana* Dain (*Everticyclammina* sp. 1 "big" sensu Ascoli, 1988), known from the Oxfordian of both regions, continued to exist and occurred for the last time on the Canadian shelf in the Kimmeridgian.

The zonal subdivision of the Kimmeridgian on the Canadian Atlantic shelf and in the East European craton is correlated according to common species of the zonal diagnostic assemblages. The Kimmeridgian calcareous benthos zone of the Canadian Atlantic shelf, *Planularia tricarinella-Epistomina mosquensis*, has common species with the East European craton lower Kimmeridgian *Lenticulina kuznetsovae-Epistomina praetatariensis* Zone (*Epistomina cognita, E. arkelli* and *Paulina marginata*) and the upper Kimmeridgian *Haplophragmium monstratum-Pseudolamarckina pseudorjasanensis* Zone (*Epistomina praereticulata* and *E. stellicostata*).

The agglutinated benthos lower Kimmeridgian *Alveosepta jaccardi* Zone and the upper Kimmeridgian *Trocholina solecensis* Zone of the Canadian shelf are correlated with the Kimmeridgian of the East European craton by the zonal index species and by the presence of *Mesoendothyra izjumiana* Dain and *Paleogaudryina varsoviensis* (Bielecka et Pozaryski) in the upper Kimmeridgian.

The Tithonian corresponds to the lower and middle Volgan and yields 11 common species. Planktonic foraminifers were not recorded.

The Tethys endemic *Anchispirocyclina lusitanica* (Egger) is characteristic of the Tithonian from the southern Canadian Atlantic shelf. It evidently appeared in the Kimmeridgian and was also found in the Tithonian of the Crimea, where it continued to exist in the

Berriasian (Kuznetsova and Gorbachik, 1985). The following species, which passed from the older sediments, occurred for the last time in the Tithonian of the Canadian shelf: *Epistomina* ex gr. *parastelligera* (Hofker), *E. ventriosa* Espitalie et Sigal, *E. stellicostata* Bielecka et Pozaryski, *E. madagascariensis* Espitalie et Sigal, *Trocholina elongata* (Leupold), *Ammobaculites coprolithiformis* (Schwager), *Paleogaudryina varsoviensis* (Biel. et Poz.), and *Planularia dofleini* Kasantsev. *Lenticulina quenstedti* (Gümbel) and *L. polonica* (Wisniowski) occurred for the last time in the lower Tithonian. *Planularia beierana* (Gümbel) passed into the Berriasian. In the East European craton, the stratigraphic intervals of some of these species are often one or two stages lower than on the Canadian shelf, and their last occurrence levels are in the Oxfordian or Kimmeridgian (Fig. 2).

The calcareous benthos *Epistomina stellicostata-E.* ex gr. *parastelligera-Epistomina ventriosa* Zone was distinguished in the Tithonian of the Canadian shelf (Ascoli, 1990). This interval corresponds to the following three lower and middle Volgan zones of the East European craton: *Lenticulina undorica-Pseudolamarckina bieleckae* (lower Volgan), *Lenticulina ponderosa-Saracenaria pravoslavlevi* (middle Volgan), and *Spirofrondicularia rhabdogonioides-Lenticula oligostegia* (middle Volgan). The lower zone, limited at the top by *Ilovayskya pseudoscythica*, with common species *Planularia dofleini* Kasantsev, *Epistomina stellicostata* Biel. et Poz., and *E. praereticulata* Mjatliuk, is correlated with the Canadian shelf sediments according to the zonal diagnostic assemblage. Zones of the middle and upper Volgan (the latter corresponds to the lower Berriasian on the Canadian shelf) have no correlative species in the zonal diagnostic assemblages.

The agglutinated benthos Tithonian *Anchispirocyclina lusitanica-Ammobaculites coprolithiformis* Zone of the Canadian shelf can be correlated according to *Anchispirocyclina lusitanica* (Egger), known from the Crimean Tithonian, and according to *Trocholina elongata* (Leupold), from the Tithonian of the Ukrainian Carpathians (Prakticheskoe Rukovodstvo ... , 1991; Kuznetsova and Gorbachik, 1985).

LOWER CRETACEOUS

We analyzed the Berriasian-Barremian interval. Such species as *Epistomina volgensis* Mjatliuk, *Verneuilinoides neocomiensis* (Mjatliuk), and *Planularia beierana* (Gümbel), passed from the Tithonian, Atlantic indicate the Tithonian-Berriasian boundary. Generally, it is a sharp boundary, because it is marked by the extinction of seven of nine species composing the upper Tithonian, zonal diagnostic assemblage of the Canadian Atlantic shelf (Fig. 3). The zonal subdivision is analyzed according to the data on the north of the East European craton (Pechora basin) and its southern margin (the Crimean-Caucasian basin) (Bystrova, 1991; Gorbachik, 1991; Kuznetsova and Gorbachik, 1985).

Seven common species were recorded in the *Berriasian-Valanginian* in both regions. The planktonic species *Favusella hoterivica* (Subbotina) most likely appeared in the upper Berriasian, on the Canadian shelf, and in the Hauterivian, in the East European craton.

The lower Berriasian sediments of the Pechora basin have no common benthic species, whereas the upper Berriasian-Valanginian yields the common species *Verneuilinoides neocomiensis* (Mjatliuk). The following forms common with the Canadian shelf deposits appeared, along with *V. neocomiensis*, in the upper Valanginian sediments of the Pechora basin (*Polyptychites polyptychus* zone): *Lenticulina eichenbergi* Bart. et Brand, *L. schreiteri* Bart. et Brand, *Planularia crepidularis* (Reuss), *Epistomina caracolla caracolla* Roemer, *E. tenuicostata* Bart. et Brand, and *E. ornata* Roemer. Among them, the first occurrence of *Planularia crepidularis*, *E. caracolla*, and *E. ornata* was recorded on the Canadian shelf in the Berriasian, whereas *Lenticulina eichenbergi*, *L. schreiteri*, and *E. tenuicostata* appeared in the Valanginian.

The Berriasian and Valanginian sediments in the southern margin of the East European craton yield the following common benthic species: *Trocholina elongata* (Leupold), with its last occurrence in the lower Berriasian; *Conorboides hofkeri* (Bart. et Brand) and *Saracenaria valanginiana* Bart. et Brand, which appeared in the upper Berriasian; and *Lenticulina busnardoi* Moullade, *L. guttata guttata* (Ten Dam), and *Epistomina ornata* (Roemer), which appeared in the Valanginian.

The zonal subdivision of the Berriasian-Valanginian, on the Canadian Atlantic shelf and in the Pechora basin, according to calcareous benthos, is similar in the upper Valanginian, where *Lenticulina eichenbergi* Bart. et Brand is an index species of the *Reophax minutissima-Lenticulina eichenbergi* Zone (Bystrova, 1991). The common species *Epistomina caracolla* (Roemer), *E. ornata* (Roemer), and *E. tenuicostata* Bart. et Brand were used on the Canadian shelf as the index species for the Hauterivian and Barremian zones, according to their last occurrence levels.

In the Crimean-Caucasian region, the planktonic foraminifer beds with *Globuligerina gulekhensis* and *G. hoterivica* were identified in the Berriasian-Valanginian and Hauterivian, respectively, for the subdivision of the beginning of the Lower Cretaceous. On the Canadian Atlantic shelf, *Favusella hoterivica* was recorded from the uppermost Berriasian to the top of the Barremian, which permitted us to distinguish the corresponding planktonic zone in this interval.

The Berriasian *Epistomina* aff. *minutereticulata-E. volgensis* Zone of the Canadian Atlantic shelf cannot be correlated by common species with the Berriasian and Valanginian calcareous benthos zones. The Crimean upper Berriasian *Conorboides hofkeri-Conorbina heteromorpha* Zone can be correlated with the Berriasian of the Canadian Atlantic shelf by the presence of *Conorboides hofkeri* (Bart. et Brand). The Crimean lower Val-

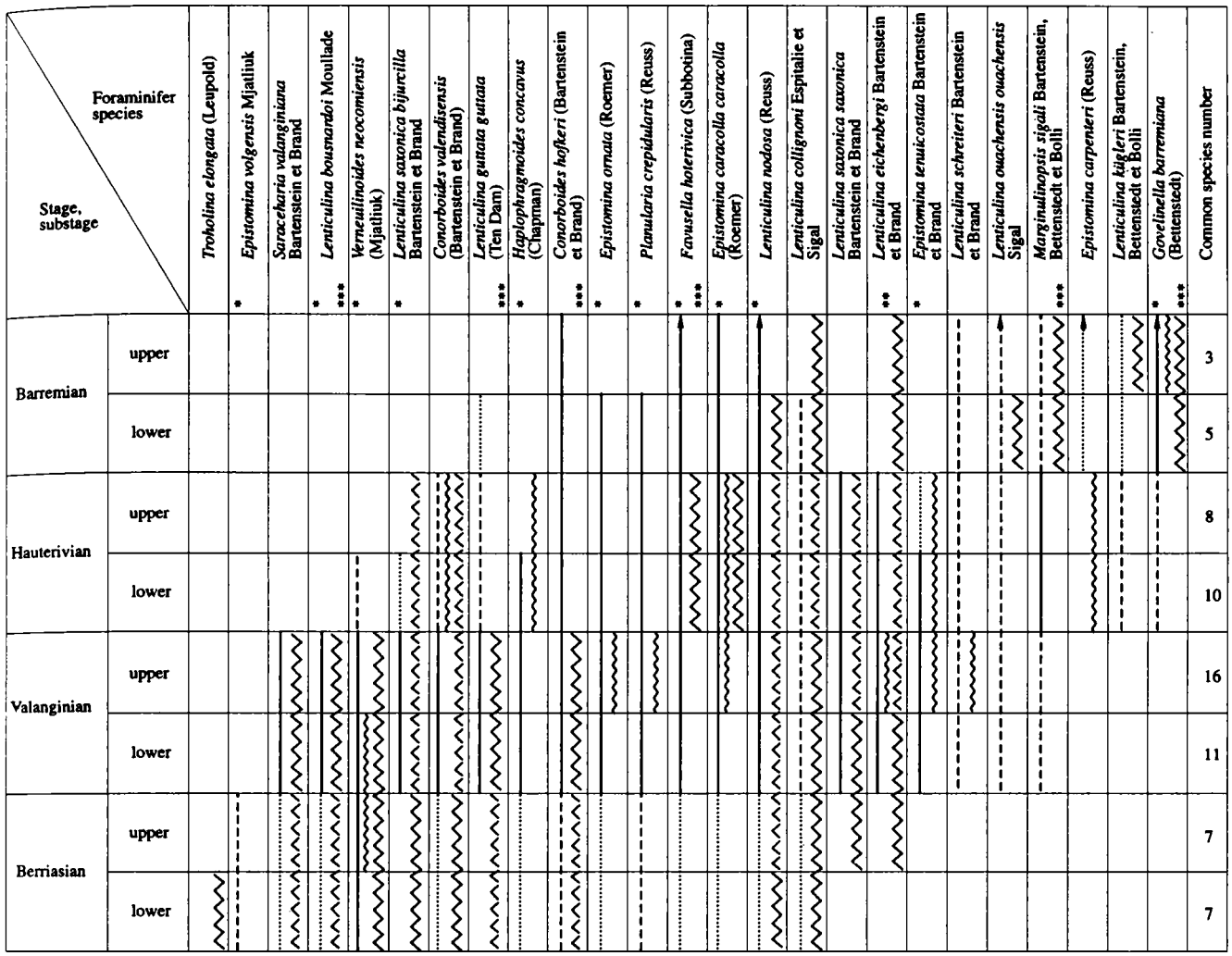

Fig. 3. *Analysis of the stratigraphic distribution of the Lower Cretaceous (Berriasian-Barremian) foraminifers on the Canadian Atlantic shelf and in the East European craton with its southern margin. For symbols see Fig. 1.*

anginian *Lenticulina busnardoi-Lenticulina guttata guttata* Zone can be correlated with the Valanginian *Lenticulina busnardoi-Lenticulina saxonica bifurcilla* Zone of the Canadian shelf by the index species.

The Berriasian-Valanginian agglutinated benthos *"Everticyclammina virguliana" plexus-Haplophragmoides concavus* Zone of the Canadian shelf has no common species with the East European craton and its southern margin sediments.

The Hauterivian sediments yield up to 11 common species. Among them, *Conorboides valendisensis* (Bart. et Brand) and *Epistomina caracolla caracolla* (Roemer) were found on both the Canadian Atlantic shelf and in the Pechora and Crimean-Caucasian basins. *Haplophragmoides concavus* (Chapman), from the Valanginian to lower Hauterivian of the Canadian Atlantic shelf, and *H.* ex gr. *concavus*, from the Hauterivian of the Pechora basin, may possibly be identical. In addition to the above-mentioned *Favusella hoterivica*, *Epistomina caracolla*, and *Conorboides valendisensis*, the Hauterivian sediments of both the Canadian Atlantic shelf and the Crimean-Caucasian region yield *Dorothia kummi* Zedler, *Lenticulina saxonica bifurcilla* Bart. et Brand, *L. saxonica saxonica* Bart. et Brand, *L. eichenbergi* Bart. et Brand, *L. nodosa* (Reuss), and *L. collignoni* Espitalie et Sigal.

The calcareous benthos *Planularia crepidularis-Epistomina tenuicostata* Zone, distinguished in the Hauterivian of the Canadian Atlantic shelf, can be correlated with the Hauterivian of the East European craton and the Crimea according to the common species mentioned above. On the Canadian shelf, a single agglutinated benthos *Verneuilinoides neocomiensis-Dorothia kummi* Zone was distinguished in the Hauterivian-Barremian. Its lower (Hauterivian) part can be correlated with the *Marginulinopsis sigali-Dorothia kummi* Zone of the lower Hauterivian of the Crimean-Caucasian region by the index species (Kuznetsova and Gorbachik, 1985; Ascoli, 1990).

The Barremian sediments yield up to seven common species of the Canadian Atlantic shelf and the East European craton southern margin. In the Pechora basin, lower Barremian sediments were not found, whereas the upper Barremian does not yield common species. In the Crimean-Caucasian basin, *Marginulinopsis sigali* Bart., Bett., et Bolli occurred only in the Barremian. The characteristic species *Gavelinella barremiana* (Bett.) continued to exist in the Aptian. The last occurrence levels of *Lenticulina nodosa* (Reuss) and *L. collignoni* Espitalie et Sigal were in the lower and in the upper Barremian, respectively. *Lenticulina ouachensis* Sigal was characteristic only of the lower Barremian. *Planularia crepidularis* (Reuss) and *Lenticulina kugleri* Bart., Bett., et Bolli continued to exist in the Aptian. In the Barremian of the Canadian Atlantic shelf, the last occurrence levels of some of the above species coincided; others differed by a half stage (Fig. 3).

The Barremian calcareous benthos zones of the Canadian Atlantic shelf and the Crimean-Caucasian region are designated by different index species. However, the Barremian *Epistomina caracolla-E. hechti-E. ornata* Zone of the Canadian shelf (characteristic of the lower part of the stage) can be correlated with the *Gavelinella barremiana-Conorotalites bartensteini* Zone of the Crimean-Caucasian region according to their common species. The Barremian subdivision of the Crimea and the Caucasus according to planktonic foraminifers is more detailed and most likely corresponds to the upper part of the *Favusella hoterivica* Zone of the Canadian Atlantic shelf.

ZONAL SUBDIVISION

The study of the stratigraphic range of the species reveals the succession of foraminifer assemblages of the same type throughout the Bajocian-Barremian interval on the Canadian Atlantic shelf and the East European craton, including its southern margin in the Crimean-Caucasian region. Foraminifer zones of both regions can be correlated according to the following criteria: (1) the same index species, (2) the common species of the zonal diagnostic assemblage, and (3) the common species of the zonal complex. The distinguished zones are substantiated by the analysis of biozones of species recorded in the zonal complex.

The difference in methods of distinguishing the zones and the choice of index species (according to the first or last occurrence) did not reveal considerable contradictions with the accepted zonal subdivision of the Middle Jurassic to Early Cretaceous in either region. Such a conclusion seems to be unbiased, because we developed the zonal subdivision in the regions separately. Differences in the stratigraphic ranges of the species, established according to our collections, undoubtedly show higher last-occurrence levels of many species on the Canadian Atlantic shelf than those in the East European craton. This fact can most likely be explained by the species' westward migration in the Boreal Atlantic basin during the opening of the North Atlantic.

PALEOECOLOGY

The zonal subdivision of the Jurassic to Lower Cretaceous sediments, according to such different ecologic groups of foraminifers as plankton and calcareous and agglutinated benthos, takes into account the association of certain foraminiferal groups with a type of sedimentation, namely, of plankton with relatively open shelf clayey-carbonate facies, of calcareous benthos with slope carbonate-clayey and reef facies, and of agglutinated benthos with shallow-water sandy-clayey facies. Study of the distribution of the distinguished ecozones in certain sections of the Canadian Atlantic shelf boreholes for the purpose of a basin analysis and facies zonation reconstruction could be an interesting object of future research.

The taxonomic composition of the Jurassic and Early Cretaceous foraminifers was the result of not only facies peculiarities but also of the paleobiogeographic (climatic) zonation. The Boreal Atlantic (nodosariid-epistominid) type of foraminiferal fauna occurred in the Jurassic to Early Cretaceous basins of the Canadian Atlantic shelf and East European craton (Basov, 1991; Grigelis, 1985). This type was characterized by a considerable taxonomic variability, a mosaic structure of populations, and the fast evolution of lenticulinids and epistominids. Certain distinctive outbreaks of the *Ammobaculites*, *Ophthalmidium*, *Trocholina*, and *Spirillina* genera were characteristic of the carbonate- and sandy-clayey shallow-water facies of the East European craton. In the deeper basins of the Canadian shelf, epistominids predominated during a continual sedimentation from the beginning of the Middle Jurassic.

The paleobiogeographic boundary of the two largest provinces (belts) of the Jurassic and Cretaceous, the Boreal and Tethyan, passed along the southern margin of the studied regions. This boundary was distinctly marked by the appearance of the Tethys species *Trocholina elongata* in the Kimmeridgian and of *Anchispirocyclina lusitanica* in the Tithonian of both regions (Ascoli, 1988). On the Canadian Atlantic margin, the latter species was recorded in the Scotian shelf boreholes (Oneida 0-25, Migrant N-20) in the Scotian (Citnalta I-59, South Griffin J-13) and Jeanne d'Arc (Bonnition H-32) basins. This boundary passed along the modern latitude 44° - 45°N.

REFERENCES

Ascoli, P., Foraminiferal and Ostracod Biostratigraphy of the Mesozoic-Cenozoic, Scotian Shelf, Atlantic Canada, *First Int. Sympos. on Benthonic Foraminifera of Continental Margins*, Maritime Sediments, Spec. Publ., 1976, no. 1, pp. 653 - 771.

Ascoli, P., Foraminiferal, Ostracode, and Calpionellid Zonation and Correlation of 42 Selected Wells from the North Atlantic Margin of North America, *Bull. Canad. Petrol. Geology*, 1990, vol. 38, no. 4, pp. 485 - 492.

Ascoli, P., Mesozoic-Cenozoic Foraminiferal, Ostracod, and Calpionellid Zonation of the North Atlantic Margin of North America: Georges Bank-Scotian Basins and Northeastern Grand Banks (Jeanne d'Arc, Carson, and Flemish Pass Basins). Biostratigraphic Correlation of 51 Wells, *Geol. Surv. Canada*, Open File, 1791, 1988, pp. 1 - 41.

Basov, V.A., Paleoecologic and Paleobiogeographic Reconstructions, *Prakticheskoe Rukovodstvo po Mikrofaune SSSR, Foraminifery Mezozoya*, vol. 5, Nedra, Leningrad, 1991, pp. 210 - 223.

Bielecka, W. and Styk, O., *Biostratygrafia Batonu i Keloweju Polnocno-Zachodniej Polski na Podstawie Otwornic i Malzoraczkow*, Prace Inst. Geologicznego, Warszawa, 1981.

Budowa Geologiczna Polski. T. III. Cz. 26. Mezozoik. Jura, Wydawnictwa Geologiczne, Warszawa, 1980, pp. 108 - 140, 291 - 326.

Bystrova, V.V., Cretaceous. Lower Series. North of the USSR European Part, *Prakticheskoe Rukovodstvo po Mikrofaune SSSR, Foraminifery Mezozoya*, vol. 5, Nedra, Leningrad, 1991, pp. 104 - 109.

Gorbachik, T.N., Cretaceous. Lower Series. South of the USSR (Planktonic Foraminifer Zones and Beds), *Prakticheskoe Rukovodstvo po Mikrofaune SSSR, Foraminifery Mezozoya*, vol. 5, Nedra, Leningrad, 1991, pp. 137 - 151.

Grigelis, A.A., *Zonal'naya Stratigrafiya Baltiiskoi Yury po Foraminiferam. Metody Analiza Foraminiferovykh Zon* (Zonal Stratigraphy of the Baltic Jurassic according to Foraminifers. Methods of the Foraminifer Zone Analysis), Nedra, Moscow, 1985.

Grigelis, A.A. and Kuznetsova, K.I., Upper Jurassic Stratigraphic Scale of the USSR according to Foraminifers, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1987, no. 5, pp. 26 - 37.

Kuznetsova, K.I. and Gorbachik, T.N., *Stratigrafiya i Foraminifery Verkhnei Yury i Nizhnego Mela Kryma* (Stratigraphy and Foraminifers of the Upper Jurassic and Lower Cretaceous of the Crimea), Nauka, Moscow, 1985.

Pazdro, O., Bathonian Globigerina of Poland, *Rocznik Polskiego Towarzystwa Geologicznego*, 1969, vol. 39, nos. 1 - 3, pp. 41 - 56.

Prakticheskoe Rukovodstvo po Mikrofaune SSSR, Foraminifery Mezozoya (Handbook on Microfauna of the USSR, Mesosoic Foraminifers), Azbel', A.Ya. and Grigelis, A.A., Eds., vol. 5, Nedra, Leningrad, 1991.

Pyatkova, D.M. and Permyakova, M.N., *Foraminifery i Ostrakody Yury Ukrainy. Paleontologicheskii Spravochnik* (Jurassic Foraminifers and Ostracodes of the Ukraine. Paleontological Handbook), Naukova Dumka, Kiev, 1978.

Srednii i Verkhnii Oksford Russkoi Platformy (Middle and Upper Oxfordian of the Russian Platform), Nauka, Leningrad, 1989.

Stratigrafiya i Korrelyatsiya Verkhnei Yury SSSR po Foraminiferam (Stratigraphy and Correlation of the USSR Upper Jurassic according to Foraminifers), Geol. Inst. USSR Acad. Sci., Moscow, 1985.

Williams, G.I., Ascoli, P., Barss, M.S., Bujak, J.P., Davies, E.H., Fensome, R.A., and Williamson, M.A., Biostratigraphy and Related Studies, Chapter 3, Geology of the Continental Margin of Eastern Canada, *Geol. Surv. Canada*, 1990, no. 2, pp. 87 - 137.

Stratigraphy and Geological Correlation, Vol. 1, No. 4, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 4, 1993, pp. 56 - 63.
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Evolution of the biosphere

Abiotic and Biotic Events in the Cenozoic Ocean

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Abstract – An analysis is given of abiotic events (fluctuations of temperature and sea level and changes in the circulation of surface and deep waters, in other paleoceanographic parameters, and in the paleogeography of the land and the seas) and of biotic events (extinction and appearance of new taxa and changes in their diversity) in the Cenozoic oceans. All events are correlated with chronologic and zonal biostratigraphic scales (zones of planktonic foraminifers and of nannoplankton). The periods of the major environmental changes and their concurrence or disagreement with biotic events are especially studied. The conclusion is drawn that the oceanic biota (planktonic and deep-water benthic foraminifers, deep-water ostracods, nannoplankton) did not respond with marked evolutionary transformations to the cardinal changes in abiotic factors.

Knowledge of the evolution of the largest ecosystems of the Cenozoic (the ocean, the epicontinental seas, the land, and the fresh-water basins) is deficient if a correlation between individual events is not established, without which it is impossible to estimate the importance and the scope of these events and their possible causes. Therefore, to begin, a calendar of abiotic and biotic events in the Cenozoic should be compiled on the basis of the most detailed chronologic and zonal biostratigraphic scale, i.e., the ocean scale.

The compilation of this calendar is based on the sequence of abiotic and biotic events in the ocean, such as the changes in the temperature, sea level, circulation patterns of surface and deep waters, other paleoceanographic parameters, and the paleogeography of the seas and the land. Among the biotic events of primary significance are the extinctions and appearances of new taxa of planktonic and deep-water benthic foraminifers, nannoplankton, radiolarians and deep-water ostracods, and other groups, and the migration of faunal and floral assemblages (Fig. 1).

In the early Paleocene, after a certain cooling that was traced only in the North Atlantic, a warming set in, and the early Paleocene climate was warm and uniform; the ocean was also warm, with low vertical and meridional gradients and slight temperature variations. The late Paleocene showed no particular changes, though there was a cooler interval at 60 Ma, both in the Atlantic and in the Southern Ocean. This short period of cooling was followed by warming, during which the temperatures in high latitudes of the Southern Hemisphere were lower than those of the same latitudes in the Northern Hemisphere, though still rather high (up to 18°C in the surface waters).

A global warming and a decrease of the pole-equator temperature gradient occurred at the Paleocene-Eocene boundary as a result of warmer waters in the high latitudes. The warming continued through most of the early Eocene. The climate was warm and uniform, with low vertical gradients. There was a slight temperature fall at the end of the early Eocene, followed by warming; the conditions in all oceans were monotonous, and meridional and vertical gradients were low. At the end of the middle Eocene, cooling, especially of the deep waters, started again and continued through the late Eocene and the beginning of the Oligocene; the temperature lowering was gradual, with occasional abrupt drops. The temperatures of the surface and bottom waters in the high latitudes of the Southern Hemisphere dropped 4 - 5°C; in the Northern Hemisphere, they were also lower. In the middle latitudes, the temperature of the surface waters (0 - 200 m) remained rather high: at 60° S, it was no less than 17°C. In the low latitudes, the cooling was insignificant (Haq *et al.*, 1977; Matthews and Poore, 1980; Keller, 1983, 1988; Kennett *et al.*, 1985; Krasheninnikov and Basov, 1986; Nikolaev, 1987; Nikolaev *et al.*, 1987; Krasheninnikov *et al.*, 1988, 1989; Barash, 1988; Barker *et al.*, 1988).

The terminal Eocene temperature decline is explained in two ways. The first and most popular hypothesis states that the cooling was a result of the ice sheet formation in the Antarctic Sea, which cooled the bottom waters (Kennett and Shackleton, 1976; Cavelier *et al.*, 1981; Keller, 1983). The ice sheet, in its turn, appeared as a result of progressing separation between

Australia and Antarctica and the thermal isolation of Antarctica (Keller, 1983). About 40 Ma ago, this process brought about fundamental changes in oceanic circulation and, consequently, the development of a psychrosphere, i.e., a lower cold layer of oceanic waters, with a high circulation rate, isolated from the upper layer or the thermosphere (Benson *et al.*, 1985). The cold Antarctic bottom waters moved to the north, and the Circum-Antarctic Current initiated (Keller, 1983). According to the other hypothesis, the cooling was caused by the injection of low-salinity waters from the previously isolated Arctic Ocean. These waters flowed over the surface waters, disturbing the heat exchange between deep and surface waters, which resulted in ice accumulation in Antarctica and in the formation of cold bottom waters (Thierstein and Berger, 1976).

At the Eocene-Oligocene boundary and at the beginning of the Oligocene, the cooling continued, the stratification of waters increased, the psychrosphere further developed, and the pole-equator gradient increased (Kennett, 1977; Keller, 1983; Benson *et al.*, 1984; Kennett *et al.*, 1985).

Therefore, the Eocene-Oligocene boundary interval is a stage in the evolution of the Paleogene ocean, when a series of global climatic events occurred, marked by transitions from the warm and probably ice-free ocean to a colder ocean with ice sheets in the Antarctic seas, larger pole-equator gradients, and more intense surface and deep water circulation.

At the beginning of the early Oligocene (35.8 Ma ago), the temperature in the deep waters of the temperate latitudes in the Southern Atlantic dropped 4 - 6°C in 10^4 - 10^5 years, apparently reflecting a temporary glaciation of Eastern Antarctica. There were two episodes of cooling (31 and 24 Ma ago), also caused by the glaciation of Antarctica, with a warmer period between them (Shackleton, 1986). The Oligocene cooling involved the surface waters as well (Keigwin, 1980; Nikolaev *et al.*, 1985), but as stated by S.D. Nikolaev (Nikolaev *et al.*, 1985), the mean global temperatures were not lower than recent temperatures because of higher temperatures in the high latitudes. At the end of the Oligocene, the ring of the deep-water Circum-Antarctic Current closed (Kennett, 1977).

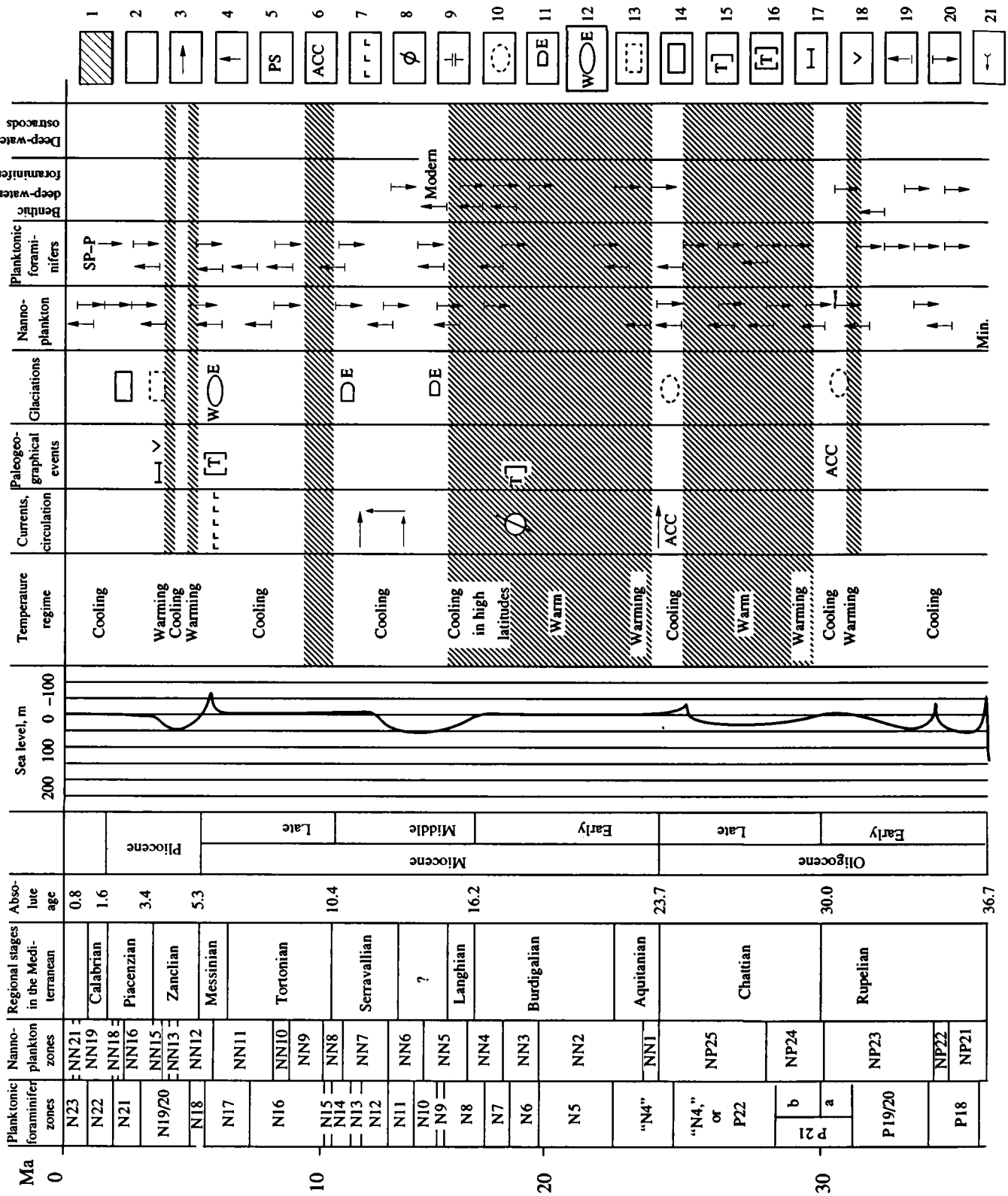
In the early Miocene (zone N4), the conditions were warm and uniform for all oceans; i.e., the temperature of surface waters in temperate and near-polar regions was higher (7 - 12°C) and, in the equatorial region, lower (23 - 26°C) than it is now. The pole-equator temperature gradients were also lower and, as a result, subtropical and temperate zones expanded to higher latitudes. At the end of the early-beginning of the middle Miocene (zone N8; 17 - 15 Ma ago), still warmer conditions set in; i.e., in the low latitudes, the temperatures grew, and the equatorial and subequatorial zones formed a continuous latitudinal band from 30° N to 30° S with surface water temperature reaching 26 - 28°C. In the middle latitudes, the climate was also warmer than it is now. At the same time, the differentiation of the climate became more pronounced: it was cooler in the lower latitudes; the temperate zone in the Southern Hemisphere moved to the north under the effect of the increasing glaciation of Antarctica and the corresponding fall of the surface water temperature in this region; and the thermal pole-equator gradient increased.

In Western Antarctica, the cooling started in the middle of the middle Miocene, whereas in Eastern Antarctica, the glaciation progressively developed and took the shape of an ice sheet, which grew until the end of the middle Miocene (zone N14). Climatic zonality became more distinct, and the temperature pole-equator gradient increased. In the equatorial zone, the surface waters remained as warm as they were earlier, but the zone itself narrowed; the surface waters in the high latitudes became colder.

In the late Miocene, the slight warming at the beginning of the interval changed to a cooling (except in low latitudes) with a further increase in the pole-equator temperature gradient, and the polar type of surface waters was thus formed for the first time. In the low latitudes, on the other hand, the period was warmer, which resulted in the expansion of warm-water zones, while the cold-water zones moved to the equator. The Southern Hemisphere was colder than the Northern Hemisphere, and the bipolar climatic zones took an asymmetrical position. The glaciation of Western Antarctica was then completed.

At the beginning of the early Pliocene, the climate, on the whole, was warmer than now; then, cooling started with intermittent short warmings, e.g., at the early-late Pliocene boundary (zone N19, 5 - 4 Ma ago), when the equatorial zone still existed; there are no contemporary analogs of such a zone. At the beginning of the late Pliocene (zone N20), this zone disappeared in the Pacific and Atlantic oceans, and it became colder in the lower latitudes, as compared with the early Pliocene. On the whole, the climate of this interval was somewhat colder than the modern climate and was colder still in the second half of the late Pliocene. The Antarctic glaciation was followed by the Arctic glaciation 3.5 Ma ago, in which ice appeared that later (2.4 Ma ago) formed into an ice sheet (Kennett and Shackleton, 1976; Kennett, 1977; Berggren and Hollister, 1977; Adams *et al.*, 1977; Clark, 1982; Rögl and Steininger, 1984; Shackleton *et al.*, 1984; Kennett *et al.*, 1985; Ivanova *et al.*, 1986; Blyum and Sokolova, 1987; Nikolaev *et al.*, 1987; Blyum *et al.*, 1988; Wright and Thunell, 1988; Barash *et al.*, 1989; Scott *et al.*, 1989).

The sea level in the Paleogene and Neogene fluctuated repeatedly. In the Paleocene, it was first at 100 - 200 m, or even somewhat below 100 m, in relation to its recent position. Later, the sea level reached about 200 or 200 - 250 m, and in the late Paleocene, it fell, for a short period, to 100 m. In the early Eocene, the level was at about 200 - 250 m, and only at the end of this interval did it fall for a short time to 100 m. After that it rose again and was at about 250 m at the beginning of the middle Eocene; then it gradually fell, and at the end of the middle Eocene, it briefly reached 100 - 150 m. Throughout the late Eocene, the sea level was at about


Ma
Absolute age
Regional stages in the Mediterranean
Nannoplankton zones
Planktonic foraminifer zones
Sea level, m
Temperature regime
Currents, circulation
Paleogeographical events
Glaciations
Nannoplankton
Planktonic foraminifers
Benthic deep-water foraminifers
Deep-water ostracods
Pliocene
Miocene
Oligocene
Early
Late
Middle
Calabrian
Piacenzian
Zanclian
Messinian
Tortonian
Serravallian
Langhian
Burdigalian
Aquitanian
Chattian
Rupelian
Cooling
Warming
Warming in high latitudes
Warm
Modern
Min.
ACC
SP-P
PS

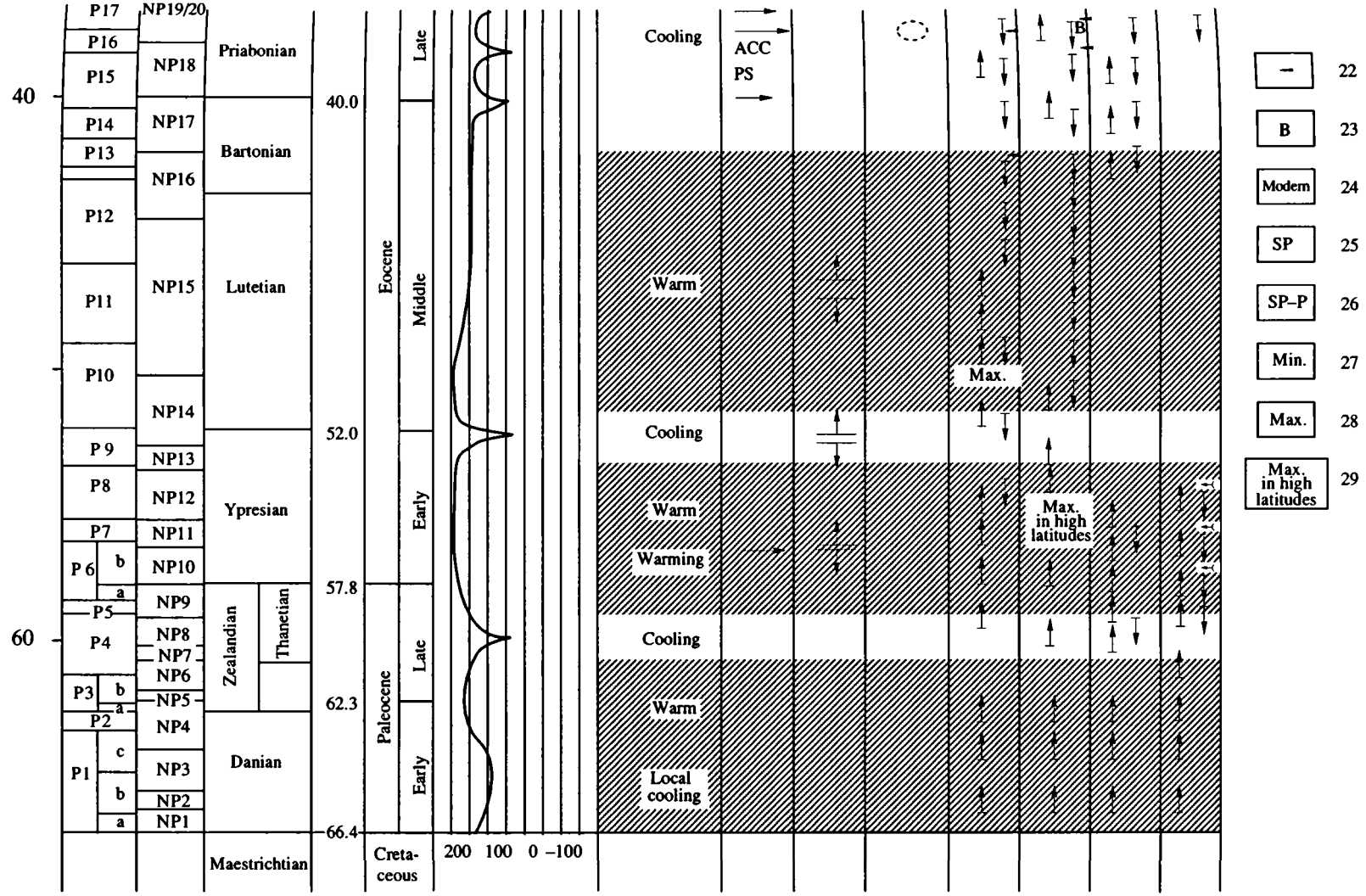


Fig. 1. *Abiotic and biotic events in the Cenozoic ocean.*

1 – warm stage; 2 – cold stage; 3 – acceleration of water circulation; 4 – development of upwellings; 5 – formation of the psychrosphere (PS); 6 – formation of the Antarctic Circumpolar Current (ACC); 7 – the Messinian "salinity crisis;" 8 – cessation of the Circum-Equatorial Current; 9 – separation of Antarctica and Australia; 10 – formation of ice in the eastern part of the Antarctic Ocean; 11 – formation of ice sheet in Eastern Antarctica; 12 – formation of ice sheet also in Western Antarctica; 13 – the appearance of ice in Arctica; 14 – formation of the Arctic ice sheet; 15 – closing of the Tethys in the east; 16 – closing of the Tethys in the west; 17 – formation of Panama Isthmus; 18 – disappearance of the Bering Bridge; 19 – the appearance of new species; 20 – extinction of species; 21 – penetration of low-latitude forms into high latitudes; 22 – immigration of high-latitude forms into low latitudes; 23 – formation of the boreal assemblage of planktonic foraminifers; 24 – formation of the modern assemblage of benthic deep-water foraminifers; 25 – formation of subpolar assemblage of planktonic foraminifers; 26 – formation of subpolar-polar assemblage of planktonic foraminifers; 27 – a minimum of nannoplankton diversity; 28 – a maximum of nannoplankton diversity; 29 – a maximum of planktonic foraminifer diversity in high latitudes.

200 m, and only during short periods of cooling did it fall to about 50 m. At the beginning of the early Oligocene, the sea level fell to –20 m; then it rose somewhat and again fell at the end of the early Oligocene (32 - 31 Ma ago) to a little below –25 m and remained low until 30 - 28 Ma ago. At the beginning of the late Oligocene, the sea level was close to that of modern time; later, it could have risen to 25 m, but at the end of the late Oligocene, it again fell for a short time to below –30 m.

The sea level in the middle Miocene was higher than it is now; it was lower at the beginning of the Miocene and in the late Miocene (12 Ma ago); during the rest of the period, it was close to the modern sea level. A significant sea level fall (by 40 - 70 m) occurred at the end of the late Miocene (about 6 Ma ago); this process closed the Mediterranean Sea and provoked the Messinian "salinity crisis." At the beginning of the Pliocene, the sea level rose and, as a consequence of that and certain tectonic events, the Mediterranean Sea again became holosaline. According to certain data, the sea level fluctuations in the Neogene were more intense (Haq *et al.*, 1987).

As a rule, the sea level drops coincided with coolings and correlated with the periods of changes in the stratification of water masses and the higher activity of bottom currents that caused erosion of sediments and hiatuses in the sediment sequence (Vail *et al.*, 1977; Cavelier *et al.*, 1981; Barton, Keller, 1982; Keller, 1983; Hallam, 1984; Miller *et al.*, 1985; Haq *et al.*, 1987).

It should be also noted that, during the late Eocene, volcanic activity was more intense. Microtectite interbeds, formed at that time (Keller *et al.*, 1983) and interpreted earlier as being a result of meteor fall, most likely relate to this phase of volcanism.

Thus, in the Cenozoic, the most important paleogeographic and paleoclimatic events, which were caused mainly by plate tectonics, are the following: the separation of Australia and Antarctica, which began at the Paleocene-Eocene boundary and which, in the Oligocene, resulted in the formation of the Circum-Antarctic Current, ice, and later, an ice sheet in Antarctica; the closing of the Tethys Ocean, first, in the east (in the early Miocene, 18 - 17 Ma ago), and then (at the end of the late Miocene), in the west; the termination of the global equatorial current and a cessation of communication between the Atlantic and Indian oceans in the Northern Hemisphere at the end of the early Miocene; the isolation of the Mediterranean Sea because of global regression and the Messinian salinity crisis at the end of the late Miocene, which coincided with the cooling in high latitudes, the appearance of an ice sheet in Western Antarctica, and the expansion of Antarctic ice over the ocean; the rise of the Isthmus of Panama and the closing of the Panama passage 3.5 Ma ago, thus cutting off the water exchange between the Atlantic and Pacific oceans in the lower latitudes; and the appearance of ice in the Arctic 3.5 Ma ago and the formation of an ice sheet there 2.4 Ma ago (Adams, 1981; Rögl and Steiningen, 1983, 1984; Barash, 1988; Barash *et al.*, 1989).

What was the reaction of oceanic biota to these abiotic events?

At the beginning of the early Paleocene, the plankton (foraminifers and nannoplankton) and benthos (deep-water foraminifers and ostracods) diversity was low, with a predominance of opportunistic and cosmopolitan organisms. At a later period, all taxa show a greater diversity and abundance or productivity (Thierstein and Berger, 1978; Benson *et al.*, 1984; Keller, 1988), which continued throughout the early Eocene. For example, in this interval, planktonic foraminifers reached their maximum diversity in the higher latitudes of both hemispheres (Krasheninnikov and Basov, 1986; Krasheninnikov *et al.*, 1988, 1989).

In the first half of the middle Eocene, nannoplankton was most diverse (zones NP14 and NP15) (Aubry, 1983). The tropical species of planktonic foraminifers dominated in low latitudes, and subpolar assemblages resembled those from the temperate zone, while the high latitudes were invaded by a few species from the low latitudes (Premoli-Silva and Boersma, 1988).

Since the second half of the middle Eocene, the biota diversity declined because of extinctions among nannoplankton and planktonic and deep-water benthic foraminifers. At that time, the nannoplankton species from the high latitudes migrated to the equator (Aubry, 1983; Miller and Katz, 1987). This process continued throughout the late Eocene, when the basic components of the tropical floras and faunas were the middle latitudinal species instead of the low latitudinal warm-water species, and even middle-high latitudinal forms dominated among the nannoplankton species. Some of the equatorial and tropical genera of planktonic and benthic (deep- and shallow-water) foraminifers disappeared, and the boreal fauna of planktonic foraminifers evolved (Cavalier *et al.*, 1981; Aubry, 1983; Miller and Katz, 1987; Premoli-Silva and Boersma, 1988).

The extinctions of planktonic and benthic organisms were coeval with the long-term cooling and gradually progressed from the second half of the middle Eocene to the beginning of the Oligocene; the most important changes in the oceanic biota occurred not at the Eocene-Oligocene boundary, but during the middle-late Eocene. For example, planktonic foraminifer diversity had been slowly reducing since the end of the early Eocene. At the end of the late Eocene, the extinction rate and the appearance of new taxa increased; this process, however, occurred, not as a single catastrophic event, but as a few episodes caused by the growing cooling. The most important changes in the deep-water benthic foraminifers also coincided with the middle-late Eocene, when the number of species reduced and the abyssal forms were substituted by the more eurybiontic taxa (Lipps, 1970; Corliss *et al.*, 1984; Berggren and Miller, 1989).

At the Eocene-Oligocene boundary, the fauna and flora assemblages underwent obvious but not catastrophic changes; the most spectacular were those in

the extratropical zone, i.e., in the high latitudes (Saunders *et al.*, 1984; Corliss *et al.*, 1984; Kennett *et al.*, 1985).

At the beginning of the early Oligocene (zone NP22), the nannoplankton diversity was at its lowest, but later, it increased somewhat (zones NP23 - 25). Since the Eocene, the high latitude assemblages migrated toward the equator (Aubry, 1983; Kennett *et al.*, 1985). At the beginning of the early Oligocene, the extinction of the typical Eocene species continued among the benthic deep-water foraminifers; later, the rate of change in this group composition was the same as it was earlier for the bathyal assemblages and much slower for the abyssal assemblages (Miller and Katz, 1987; Berggren and Miller, 1989). The process of the extinction of planktonic foraminifers continued, and the cosmopolitan and cold-water species became prevalent (Keller, 1983; Corliss *et al.*, 1984). As a result of all these changes, the warm-water assemblages were gradually substituted by the more cold-water ones of impoverished composition, with numerous cosmopolitan species.

In the early Miocene (zone N4), among the benthic deep-water foraminifers, the Paleogene species were being replaced by modern forms (Miller and Katz, 1987). The changes in planktonic groups throughout the Miocene were evidenced only in the extinction and the appearance of a few species (Blyum and Sokolova, 1987; Nikolaev *et al.*, 1987; Barash, 1988; Blyum *et al.*, 1988; Wright and Thunell, 1988; Barash *et al.*, 1989).

In the Pliocene, the character of this process remained unchanged; i.e., in the interval of zone N20, the equatorial assemblage of planktonic foraminifers disappeared, and during the time of zone N21 (2.4 Ma ago), a subpolar-polar assemblage was formed. This process was accompanied by the extinction of some species and the appearance of a few new ones (Wright and Thunell, 1988; Barash *et al.*, 1989).

A comparison of abiotic and biotic events in the Paleogene-Neogene ocean, in order to reveal concurrences or discrepancies between them, showed that the most abrupt changes in climate, oceanic circulation, and other abiotic factors, caused mainly by plate movements and by other tectonic activity, took place at the end of the Eocene and the beginning of the Oligocene. Temperature was, apparently, the principal factor that brought about the changes in the biota of the ocean. The oceanic biota, however, did not respond to these sharp environmental changes by essential evolutionary transformations. During these periods, a redistribution of taxa and assemblages mainly took place, accompanied by the extinction and appearance of a few species and of still fewer genera.

REFERENCES

Adams, C.G., An Outline of Tertiary Palaeogeography, *The Evolving Earth*, Cocks, L.R.M., Ed., Brit. Mus. Nat. Hist., Cambridge, Univ. Press, London, 1981, pp. 221 - 235.

Adams, C.G., Benson, R.H., Kidd, R.B., Ryan, W.B.F., and Wright R.C., The Messinian Salinity Crisis and Evidence of Late Miocene Eustatic Changes in the World Ocean, *Nature*, 1977, vol. 269, no. 5627, pp. 383 - 386.

Aubry, M.-P., Late Eocene and Early Oligocene Calcareous Nannoplankton Biostratigraphy and Biogeography, *Bull. Am. Assoc. Petrol. Geologists*, 1983, vol. 67, p. 415.

Barash, M.S., Paleoecological Aspects of Paleoceanic Evolution of Northern Atlantic and of the Adjacent Area of the Arctic Basin, *Chetvertichnaya Paleoekologiya i Paleogeografiya Severnykh Morei*, Nauka, Moscow, 1988, pp. 53 - 67.

Barash, M.S., Blyum, N.S., Burmistrova, I.I., *et al.*, *Neogen-Chetvertichnaya Paleookeanologiya po Mikropaleontologicheskim Dannym* (The Neogene-Quaternary Paleoceanology From Micropaleontological Data), Barash, M.S., Ed., Nauka, Moscow, 1989.

Barker, P.F. and Kennett, J.P., Weddell Sea Paleoceanography: Preliminary Results of ODP Leg 113, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 1988, vol. 67, nos. 1 - 2, pp. 75 - 102.

Barron, J.A. and Keller G., Widespread Miocene Deep-Sea Hiatuses; Coincidence with Periods of Global Cooling, *Geology*, 1982, vol. 10, pp. 577 - 581.

Benson, R.H., Chapman, R.E., and Deek, L.T., Paleoceanographic Events and Deep-Sea Ostracods, *Science*, 1984, vol. 224, no. 4655, pp. 1334 - 1336.

Berggren, W.A. and Hollister, C.D., Plate Tectonics and Pale ocirculation Commotion in the Ocean, *Tectonophys.*, 1977, vol. 38, nos. 1 - 2, pp. 11 - 48.

Berggren, W.A. and Miller, K.G., Cenozoic Bathyal and Abyssal Calcareous Benthic Foraminiferal Zonation, *Micropaleontol.*, 1989, vol. 35, no. 4, pp. 308 - 320.

Blyum, N.S., Ivanova, E.V., and Os'kina, N.S., Climatic Zonality of the Atlantic, Indian and Pacific Oceans in the Middle and Late Miocene, *Dokl. Akad. Nauk SSSR*, 1988, vol. 301, no. 5, pp. 1167 - 1170.

Blyum, N.S. and Sokolova, E.A., Climatic Zonality of the Pacific in the Miocene from Analysis of Planktonic Foraminifer Assemblages, *Byul. MOIP, Otd. Geol.*, 1987, vol. 62, no. 6, pp. 116 - 130.

Cavelier, C., Chateauneuf, J.J., Pomerol, Ch., Rabussier, D., Renard, M., and Vergnaud-Grazzini, C., The Geological Events at Eocene-Oligocene Boundary, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 1981, vol. 36, nos. 3 - 4, pp. 223 - 248.

Clark, D.L., The Arctic Ocean and Post-Jurassic Paleoclimatology, *Climate in Earth History*, Nat. Acad. Press, Wash., 1982, p. 133 - 138.

Corliss, B.H., Aubry, M.P., Berggren, W.A., Fenner, J.M., Keigwin, L.D., and Keller, G., The Eocene/Oligocene Boundary Event in the Deep Sea, *Science*, Wash., 1984, vol. 226, no. 4676, p. 806 - 810.

Hallam, A., Pre-Quaternary Sea-Level Changes, *Ann. Rev. Earth and Planet. Sci.*, 1984, vol. 12, pp. 205 - 243.

Haq, B.U., Hardenbol, J., and Vail, P.R., Chronology of Fluctuating Sea Level Since the Triassic, *Science*, 1987, vol. 235, no. 4793, pp. 1156 - 1166.

Haq, B.U., Premoli-Silva, L., and Lohman, G.P., Calcareous Plankton Paleobiogeographic Evidence for Major Climatic Fluctuations in the Early Cenozoic Atlantic Ocean, *J. Geophys. Res.*, 1977, vol. 82, no. 27, pp. 3861 - 3876.

Ivanova, E.V., Os'kina, N.S., and Blyum, N.S., Climatic Zonality of the Atlantic, Indian and Pacific Oceans in the Early Miocene, *Dokl. Akad. Nauk SSSR*, 1986, vol. 288, no. 5, pp. 1185 - 1188.

Keigwin, L.D., Paleoceanographic Change in the Pacific at the Eocene-Oligocene Boundary, *Nature*, 1980, vol. 287, no. 5784, pp. 722 - 725.

Keller, G., Biotic Turnover in Benthic Foraminifers across the Cretaceous-Tertiary Boundary at El Kef, Tunisia, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 1988, vol. 66, nos. 3 - 4, pp. 153 - 172.

Keller, G., Paleoclimatic Analyses of Middle Eocene through Oligocene Planktic Foraminiferal Faunas, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 1983, vol. 43, nos. 1 - 2, pp. 73 - 94.

Kennett, J.P., Cenozoic Evolution of Antarctic Glaciation, the Circum-Antarctic Ocean, and Their Impact on Global Paleoceanography, *J. Geophys. Res.*, 1977, vol. 82, no. 27, pp. 3843 - 3860.

Kennett, J.P., von der Borch, C., Baker, P.A., *et al.*, Palaeotectonic Implications of Increased Late Eocene-Early Oligocene Volcanism from South Pacific DSDP Sites, *Nature*, 1985, vol. 316, no. 6028, pp. 507 - 511.

Kennett, J.P. and Shackleton, N.J., Oxygen Isotope Evidence for the Development of the Psychrosphere 38 Ma Ago, *Nature*, 1976, no. 260, pp. 513 - 515.

Krasheninnikov, V.A. and Basov, I.A., Cenozoic Stratigraphy of the Southern Ocean, *Trudy GIN Akad. Nauk SSSR*, 1986, issue 410.

Krasheninnikov, V.A., Serova, M.Ya., and Basov, I.A., The Paleogene Zonal Scale of Planktonic Foraminifers and Its Significance for Stratigraphic Schemes of Paleogene in High Latitudes, *Kainozoi Dal'nego Vostoka, DVO Akad. Nauk SSSR*, Vladivostok, 1989, pp. 4 - 22.

Krasheninnikov, V.A., Serova, M.Ya. and Basov, I.A., Stratigraphy and Planktonic Foraminifers of the Paleogene in the Pacific High Latitudes, *Trudy GIN Akad. Nauk SSSR*, 1988, issue 429.

Lipps, J.H., Plankton Evolution, *Evolution*, 1970, vol. 24, no. 1, pp. 1 - 22.

Matthews, R.K. and Poore, R.Z., Tertiary $\delta^{18}O$ Record and Glacial-Eustatic Sea-Level Fluctuations, *Geology*, 1980, vol. 8, no. 10, pp. 501 - 504.

Miller, K.G. and Katz, M.E., Oligocene to Miocene Benthic Foraminiferal and Abyssal Circulation Changes in the North Atlantic, *Micropaleontol.*, 1987, vol. 33, no. 2, pp. 97 - 149.

Miller, K.G., Mountain, G.S. and Tucholke, B.E., Oligocene Glacio-Eustasy and Erosion on the Margins of the North Atlantic, *Geology*, 1985, vol. 13, no. 1, pp. 10 - 13.

Nikolaev, S.D., Blyum, N.S., and Nikolaev, V.I., An Attempt of Reconstruction of the Thermal Structure of Surface Water Masses in the Oligocene Ocean, Abstracts of Papers, *Vtoraya Tikhookeanskaya Shkola po Morskoi Geologii i Geofizike*, (Second Pacific Workshop on Marine Geology and Geophysics), Yuzhno-Sakhalinsk, 1985, pp. 177 - 179.

Nikolaev, S.D., Blyum, N.S., and Nikolaev, V.I., Temperature Changes in the Surface Water Masses of the World Ocean in the Eocene, *Dokl. Akad. Nauk SSSR*, 1987, vol. 297, no. 4, pp. 967 - 969.

Nikolaev, S.D., Blyum, N.S., Nikolaev, V.I., and Lisitsyn, A.P., Temperature Changes in the Surface Water Masses of the World Ocean in the Miocene, *Dokl. Akad. Nauk SSSR*, 1987, vol. 292, no. 2, pp. 464 - 467.

Premoli-Silva, I. and Boersma, A., Atlantic Eocene Planktonic Foraminiferal Historical Biogeography and Paleohydrographic Indices, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 1988 vol. 67, nos. 3 - 4, pp. 315 - 356.

Rögl, F. and Steininger, F.F., Neogene Paratethys, Mediterranean and Indo-Pacific Seaways, *Fossils and Climate*, Brenchley, P.I., Ed., Wiley and Sons, Chichester, New York, etc., 1984, pp. 171 - 200.

Rögl, F. and Steininger, F.F., Vom Zerfall der Tethys zu Mediterran und Paratethys. Die Neogene Paläogeography und Palinspastik des Zirkummediterranen Raumes. *Ann. Naturhist. Mus.*, Wien, 1983, vol. 85, ser. A, pp. 135 - 163.

Saunders, J.B., Bernoulli, D., Müller-Merz, E., *et al.*, Stratigraphy of the Late Middle Eocene to Early Oligocene in the Bath Cliff Section, Barbados, West Indies, *Micropaleontol.*, 1984, vol. 30, no. 4, pp. 390 - 425.

Scott, D.B., Mudie, P.J., Baki, V., MacKinnon, K.D, and Cole, F.E., Biostratigraphy and Late Cenozoic Paleoceanography of the Arctic Ocean: Foraminiferal, Lithostratigraphic, and Isotopic Evidence, *Bull. Geol. Soc. Amer.*, 1989, vol. 101, pp. 260 - 277.

Shackleton, N.J., Paleogene Stable Isotope Events, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 1986, vol. 57, nos. 1 - 2, pp. 92 - 102.

Shackleton, N.J., Backman, J., Zimmerman, H., *et al.*, Oxygen Isotope Calibration on the Onset of Ice-Rafting and History of Glaciation in the North Atlantic Region, *Nature*, 1984, vol. 307, no. 5952, pp. 620 - 623.

Thierstein, H.R. and Berger, W.H., Injection Events in Ocean History, *Nature*, 1978, vol. 276, no. 5687, pp. 461 - 466.

Vail, P.R., Mitchum, R.M. Jr., and Thompson, S., III, Seismic Stratigraphy and Global Changes of Sea Level, Part 4: Global Cycles of Relative Changes of Sea Level, *Mem. Amer. Assoc. Petrol. Geol.*, 1978, vol. 26, pp. 83 - 97.

Wright, J.D. and Thunell, R.C., Neogene Planktonic Foraminiferal Biogeography and Paleoceanography of the Indian Ocean, *Micropaleontol.*, 1988, vol. 34, no. 3, pp. 193 - 216.

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Evolution of the biosphere

Late Maestrichtian Flora from the Emima-Il'naivaam Interfluve, the Northeastern Koryak Highland, and Its Stratigraphic Significance

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Abstract – The Koryak taphoflora, consisting of 33 species of horsetails, ferns, gymnosperms, and angiosperms, was recorded from several localities of the upper Koryak Subformation in the Emima-Il'naivaam interfluve. Occurrences of marine mollusks in the deposits underlying and overlying the flora-bearing beds indicate the middle-late Maestrichtian age of the taphoflora. Paleobotanical data (correlation with fossil floras from the Rarytkin Ridge and Lake Pekul'neiskoe) allows specifying its age as the late Maestrichtian.

Rich and diverse Late Cretaceous floras from the northeastern Russia have long attracted the attention of paleobotanists. Many sections in this region show an alternation of flora-bearing beds with marine deposits containing abundant fragments of ammonites and inocerams. This affords a more precise definition of the stratigraphic position of these beds. The flora-bearing deposits of the Koryak Formation belong to such sections. The fossil plants from these deposits existed near the Cretaceous-Paleogene boundary, i.e., a limit that coincided with the greatest ecological crisis in the Earth's history. This flora represents a final stage of the Cretaceous florigenesis in northeastern Asia. Detailed data on this flora, including a list of species, was provided by Vasilevskaya and Abramova (1974). However, as a whole, the flora has not been studied adequately, and an attempt has been made to restudy it.

The Koryak Formation is distributed in two areas of the northeastern Koryak Highland: on the coast of the Ugol'naya Bay and in the Emima-Il'naivaam interfluve (Fig. 1). In the first area, the Koryak Formation lies conformably on the flora-bearing deposits of the Barykov Formation and is composed of tuffaceous sandstones, tuffites, siltstones, and claystones containing the remains of marine mollusks and vertebrates of late Campanian and Maestrichtian age. The uppermost strata of this section are coal-bearings deposits ("Podsopochnyi" coal bed, according to M.I. Bushuev). Rare plant remains, *Sphenopteris* sp. (*S.* ex gr. *onkilonica* Krysht.), *Thuja* sp., *Taxites* cf. *olrikii* Heer, and *Phyllites* sp. (identified by A.F. Efimova), were found at their top (Volobueva and Terekhova, 1974). The Koryak Formation is overlain conformably or with a stratigraphic hiatus by the Chukotka Formation.

In the area of the Amaam Lagoon and the Il'naivaam River (Fig. 1), the Koryak Formation stretches in a sublatitudinal direction along the northern margin of the Amaam depression. It is also distributed on the sea coast, south and north of the Amaam Lagoon, and on the right bank of the Emima River flowing into it (Fig. 2). The formation, 700 - 1000 m thick, overlies conformably the Barykov Formation and can be divided into three parts (Volobueva and Terekhova, 1974). The lower subformation, 300 - 500 m thick, is composed mainly of volcanimictic and polymictic sandstones, with a subordinate amount of siltstones, tuffstones, and acid tuffs. These deposits contain abundant remains of ammonites and inocerams, which suggest assigning the subformation to the lower half of the upper Campanian, except for its base (Volobueva and Terekhova, 1974). The middle subformation, 300 - 500 m thick, shows an increasing amount of tuffaceous material: it is represented mainly by acid tuffs, tuffites, tuffstones, and tuffaceous siltstones. There are rare *Inoceramus* ex gr. *balticus* Boehm., *Parallelodon sachalinensis* Schmidt, *Thracia* sp., *Hypophylloceras* sp. indet., *Gaudryceras* sp., *Canadoceras newberryanum* (Meck), *C. kossmati* Mat., and *C.* aff. *multicostatum* Mat., in these rocks, suggesting assigning the subformation to the upper half of the upper Campanian and, possibly, to the base of the Maestrichtian (Volobueva and Terekhova, 1974). The upper subformation, lying conformably on the middle one, is represented by continental deposits: sandstones, tuffstones, tuffites, and siltstones, with interbeds of conglomerates, car-

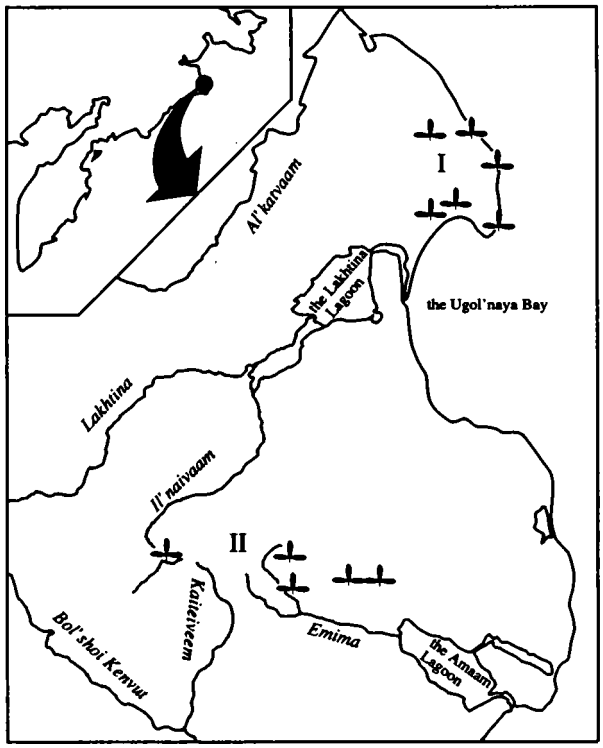

Fig. 1. *Distribution of the Upper Cretaceous flora-bearing deposits in the northeastern Koryak Upland.*

I – the Ugol'naya Bay; II – the Emima-Il'naivaam interfluve.

bonaceous siltstones, and coals. There are abundant plant remains in these deposits. The thickness of the upper subformation is 80 - 200 m.

The Amaam Formation fills the central part of the Amaam depression and lies unconformably on the Koryak Formation. The Amaam Formation, about 1500 m thick, is constituted by conglomerates, granule gravelstones, sandstones, and siltstones, with a conglomerate bed, 10 - 30 m thick, at the base. The formation was subdivided into three parts. Volobueva and Terekhova (1974) discovered *Glycymeris* (*Glycymeris*) *amaamensis* Volobueva, *Crassatella* cf. *unioides* (Stanton), *Herella* sp., and *Turritella* aff. *pachecoensis* Stanton in the sandstones and siltstones of the lower subformation above the basal conglomerate. This subformation is considered to be of the Danian-Paleocene age, and the Amaam Formation is dated as the Danian to late Eocene and, possibly, the earliest Oligocene (Volobueva and Krasnyi, 1979; Volobueva and Terekhova, 1974).

In 1987 and 1990 (in 1987, working with L.B. Golovneva, from the Botanical Institute, Russian Academy of Sciences), I examined some sections of the Koryak and Amaam formations in the Emima-Il'naivaam interfluve and collected plant remains from five localities of the upper Koryak Subformation (Fig. 2). The most complete and well-exposed section of the subformation was described from the basin of the Il'naivaam River, near the mouth of Vodosbornyi Creek (Fig. 2, Locality 1). There (Fig. 3), greenish- and yellowish-grey tuffstones and sandstones of the middle Koryak Subformation are overlain by the following members (from below upward).

M e m b e r 1. Interbedded fine-grained sandstones, tuffstones, siltstones (predominant), and claystones. The rocks are light to dark-grey, rubbly, with a shell-like joint. There is an interbed of carbonaceous siltstone in the upper third (in thickness). The thickness of the individual interbeds varies from a few centimeters to a few tens of centimeters. The member is 18 - 20 m thick.

A boundary between this member and the underlying middle Koryak Subformation is drawn, rather tentatively, according to the first occurrence of carbonaceous siltstones. In general, the contacting deposits are very similar lithologically and no hiatus can be observed.

M e m b e r 2. Fine- and medium-grained sandstones and siltstones with an admixture of tuffaceous material, gray, greenish- and yellowish-gray, firm, and well-bedded. Abundant, well-preserved plant remains were collected from the lower part of the member (in a bed about 40 - 50 cm thick) (Table 1, Locality 1). *Metasequoia occidentalis* (Newb.) Chaney, *Microconium beringianum* Golovn, *Platanus raynoldsii* Newberry, *Corylus beringiana* (Krysht.) Golovn., and *Trochodendroides bidentata* Vassilevsk. et Golovn are predominant. Less frequent are *Onoclea hesperia* R.W. Brown, *Trochodendroides* ex gr. *arctica* (Heer) Berry, *Trochodendrocarpus arcticus* (Heer) Krysht., and *Haemanthophyllum cordatum* Golovn. Imprints of other plants are sparse. G.M. Brattseva has studied a spore-pollen assemblage from the flora-bearing deposits, which appeared to be very poor: a few fern spores of the Polypodiaceae family and conifer pollen of the Taxodiaceae family (possibly, *Metasequoia*), as well as fragments of pollen grains of the Pinaceae family. Interbeds of carbonaceous siltstones are observed in the upper part of the member. The thickness of the member is 18 - 21 m.

M e m b e r 3. Sandstones and tuffstones, from fine- to coarse-grained, gray-green, thin-bedded, with large (up to 20 cm across) carbonate concretions. The thickness is 4 - 4.5 m.

M e m b e r 4. Alternation of green-gray, light-gray, and brown medium- and coarse- to very coarse-grained sandstones and tuffstones, containing plant detritus, poorly preserved leaf remains (unidentifiable), and concretions up to 10 cm across, with splintered siltstones and claystones; in places, carbonaceous or ferruginous. There is a bed, 0.6 m thick, permeated with burrows of mud-eaters in the upper part of the member. The thickness of the member is 65 - 70 m.

M e m b e r 5. Sandstones, medium- to very coarse-grained, gray, yellowish-gray, ferruginous, flaggy, with interbeds of siltstones and carbonaceous siltstones, judging from debris. In places (particularly,

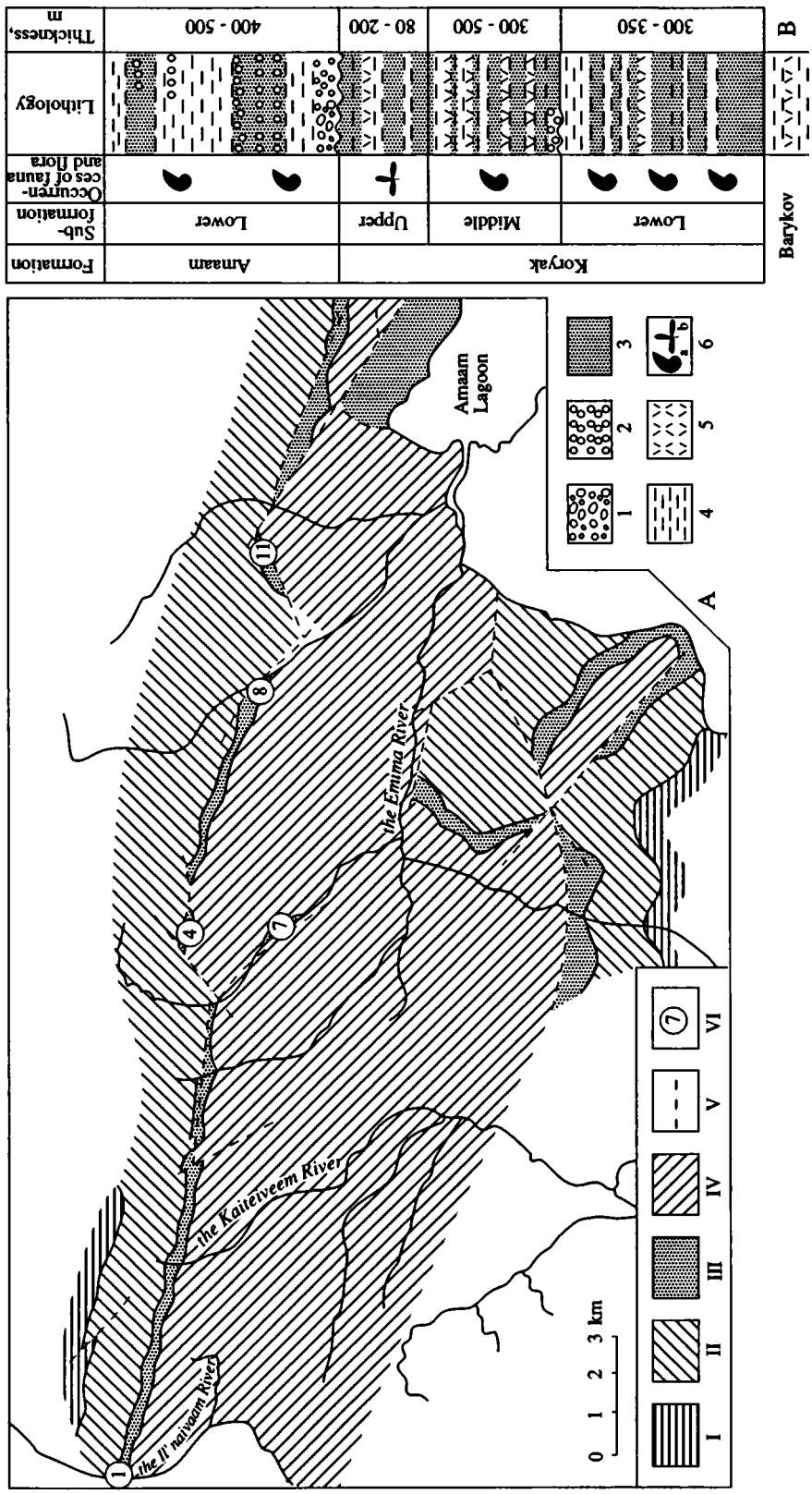

Fig. 2. *Flora-bearing deposits in the Emima-Il'naivaam Interfluve (according to Volobueva and Terekhova, with supplements).*
A – Scheme of distribution of the deposits; B – Stratigraphic column. I – Barykov Formation; II – lower and middle Koryak Subformations; III – upper Koryak Subformation; IV – Chukotka Formation; V – faults; VI – localities of the fossil flora (see Table 1).
1 – Conglomerate; 2 – granule gravelstone; 3 – sandstone; 4 – silstone and carbonaceous silstone; 5 – tuffs; 6 – occurences of the fossil fauna (a) and flora (b).

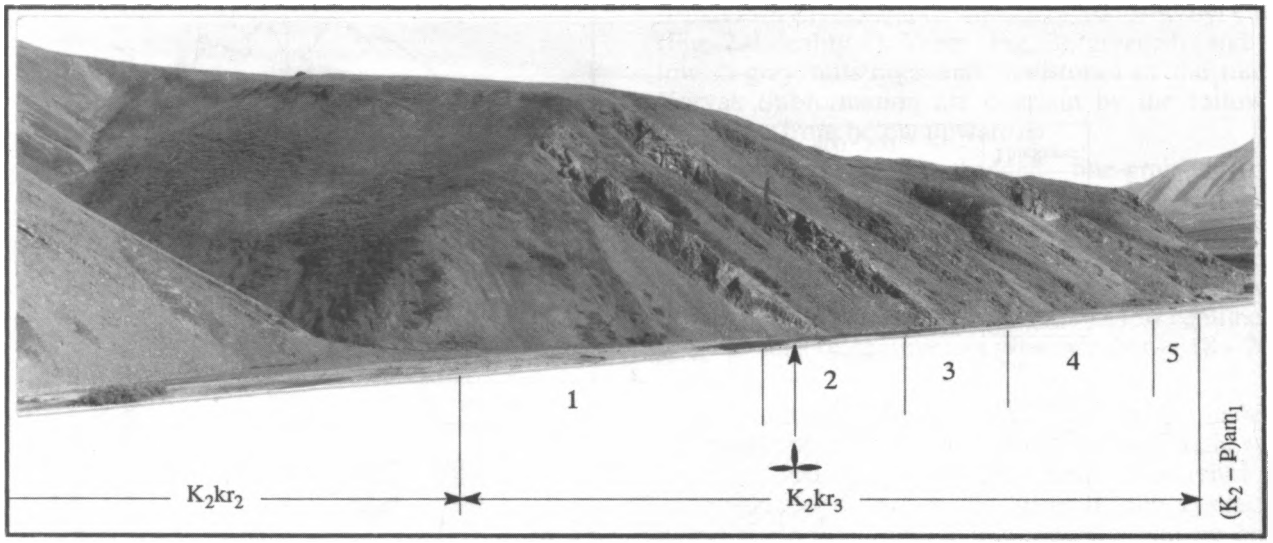

Fig. 3. *Section of the Koryak flora-bearing deposits on the Il'naivaam River (Locality 1).*
K_2kr_2 – Middle Koryak Subformation; K_2kr_3 – upper Koryak Subformation; $(K_2 - P)am_1$ – lowerAmaam Subformation; Numerals denote members of the upper Koryak Subformation.

in the upper part of the member), the sandstones contain pebbles. The thickness is about 10 m.

The total thickness of the upper Koryak Subformation is 115 - 125 m.

It is overlain by compact conglomerates with pebbles of various size and interbeds of granule gravelstones and coarse- to very coarse-grained sandstones, which form the base of the Amaam Formation. A hiatus between the Koryak and Amaam Formation is rather small, if there is one: the coarse-to very coarse-grained sandstones of the Member 5 contain, in some places, a considerable amount of pebbles (nearly conglomerates), and the conglomerates of the base of the Amaam Formation lie on the surface of the Member 5, leveled parallel to the beds. No significant traces of erosion could be observed.

In addition to the locality on the Il'naivaam River, plant fossils have been collected from four other localities of the upper Koryak Subformation at the northern side of the Amaam depression (Fig. 2, Table 1).

Locality 4 is on the left tributary of the Gravelitovyi Creek, 1.7 km upstream from its mouth. The upper Koryak Subformation overlies here the sedimentary tuffaceous strata of the middle Koryak Subformation and is represented by an alternation of sandstones and tuffstones with siltstones and carbonaceous siltstones. Plant remains with predominant *Microconium beringianum* Golovn., *Trochodendroides bidentata* Vassilevsk. et Golovn., and *Corylus beringiana* (Krysht.) Golovn. were collected from the lower part of the subformation, at a level of 5 - 8 m from its base (Table 1).

Locality 7 is at the mouth of Skalistyi Creek, the left lower tributary of Gravelitovyi Creek. Gray, medium- to coarse-grained, compact sandstones with interbeds of carbonaceous siltstones are exposed on the slope. These deposits contain well-preserved plant remains (Table 1) and belong, according to Volobueva and Terekhova, to the upper member of the upper Koryak Subformation. *Metasequoia occidentalis* (Newb.) Chaney, *Trochodendroides bidentata* Vassilevsk. et Golovn., and *Corylus beringiana* (Krysht.) Golovn. are most frequent. Imprints of *Trochodendrocarpus arcticus* (Heer) Krysht., *Cissites volobuevae* Herman, and *Rarytkinia* aff. *terechovae* Vassilevsk. et Golovn. are also abundant there.

In addition to these collections, M.A. Akhmet'ev and I studied a few samples collected in 1985 by V.E. Arkhipov and M.K. Ivanov (Moscow State University) from this locality. Among them, *Cephalotaxopsis* cf. *magnifolia* Font., *Metasequoia occidentalis* (Newb.) Chaney, *Corylus* cf. *jeliseevii* Krysht., *Grewiopsis* sp., and *Cissites* sp. were identified.

Locality 8 is the washout at the right bank of the Severnaya River, 8 km from its mouth. Plant remains (Table 1) were found in a member of siltstones and fine-grained sandstones that is gray and hard, and forms sharp-edged flags. Imprints of *Glyptostrobus nordenskioldii* (Heer) R.W. Brown, *Platanus raynoldsii* Newberry, and *Corylus beringiana* (Krysht.) Golovn. are predominant.

Locality 11 is at the southeastern slope of Mt. Medvezh'ya, on the large, left tributary of Medvezhii Creek. There is an exposure of gray, medium- and fine-grained sandstones on the right bank of the creek containing abundant well-preserved plant remains (Table 1). The most frequent are *Metasequoia occidentalis* (Newb.) Chaney, *Trochodendroides bidentata* Vassilevsk. et Golovn., *Platanus raynoldsii* Newberry, and *Celastrinites septentrionalis* (Krysht.) Golovn.

Vasilevskaya and Abramova studied plant remains collected by Volobueva and Terekhova from this loca-

lity. A species list presented by them (Vasilevskaya and Abramova, 1974) includes 30 species and is almost identical with that discussed above (Table 1). However, I have not found *Woodwardia, Sterculia, Vitis, Viburnum*, and several other plants (represented mostly by new taxa) mentioned by the these authors.

In addition to the plant fossils, remains of the early Maestrichtian mollusks *Patagiosites alaskensis* Jones, *Inoceramus piloensis* Sok., and *Inoceramus* sp. (identified by V.P. Pokhialainen) were found in the tuffstones of the middle Koryak Subformation, about 80 - 90 m below its top, on the left bank of the Emima River, in the upper course of the large right tributary of Bezymyannyi Creek.

The studied localities are confined both to the lower (Localities 1, 4) and to the upper (Locality 7) parts of the upper Koryak Subformation. A distance between the extreme flora-bearing sites is nearly 25 km in a straight line (Fig. 2). But the close taxonomic similarity of the plant fossils from these localities (Table 1) permits us to consider them as a single floral assemblage (taphoflora).

As a result of studying the new collections and of revising the samples collected earlier by Volobueva and Terekhova, 33 species have been identified from the Koryak taphoflora (Table 1). These include horsetails (one species), ferns (three species), conifers (ten species), Ginkgoals (one species), and angiosperms (18 species). The horsetails are represented by the single species *Equisetum arcticum* Heer. The ferns are not numerous. *Onoclea hesperia* R.W. Brown, characteristic of this flora, is not abundant in the localities. Occurrences of other ferns are solitary and fragmentary. Remains of *Ginkgo* ex gr. *adiantoides* (Unger) Heer are also sparse.

Though rather diverse taxonomically, conifers seldom predominate in the localities. The most abundant are remains of *Metasequoia occidentalis* (Newb.) Chaney and *Microconium beringianum* Golovn. Other conifers are not numerous. The Pinaceae family is represented by imprints of winged seeds and needles belonging to *Pityospermum* sp. and *Pityophyllum* sp. Seed scales of the *Pseudolarix* genus were found as well. Representatives of this family may have grown at some distance from the burial places of the plant remains. Shoots with scale-like, spirally arranged leaves were referred to *Araucarites conophyllus* Golovn. Their morphology is similar to that of *Araucarites*, of the Tsagayan flora from the Amur Region, and of the Late Cretaceous flora from the Vilyui basin. Remains of *Sequoia* from the Koryak Formation are characterized by a specific morphology of shoots with broad, flat, decurrent leaves. Later, they may be assigned to a new species. *Elatocladus talensis* Golovn., characterized by shoots with large, linear, spirally arranged leaves (Golovneva and Herman, 1992), cannot be related to any recent genus or family.

Most of the angiosperms are represented by woody dicotyledons. Monocotyledons are represented by

Table 1. Occurrences of the plant remains in the localities of the upper Koryak Subformation in the Emima-Il'naivaam interfluve (Golovneva and Herman's identifications)

Species	Locality				
	1	4	7	8	11
1. *Equisetum arcticum* Heer				+	
2. *Asplenium dicksonianum* Heer			+		+
3. *Onoclea hesperia* R.W. Brown	+		+	+	
4. *Cladophlebis* sp.	+				
5. *Ginkgo* ex gr. *adiantoides* (Unger) Heer					+
6. *Pseudolarix arctica* Kryshtofovich				+	
7. *Pityophyllum* sp.			+	+	+
8. *Pityospermum* sp.				+	
9. *Elatocladus talensis* Golovn.			+	+	
10. *Araucarites conophyllus* Golovn.		+			
11. *Sequoia amaamensis* Abramova et Golovn.		+			+
12. *Metasequoia occidentalis* (Newb.) Chaney	!	+	!	+	!
13. *Metasequoia* sp. (cone)	+		+		
14. *Glyptostrobus nordenskioldii* (Heer) R.W. Brown	+	+		!	
15. *Microconium beringianum* Golovn.	!	!	+	+	
16. *Haemanthophyllum cordatum* Golovn.	+				
17. *Trochodendroides arctica* (Heer) Berry	+	!			+
18. *T. bidentata* Vassilevsk. et Golovn.	!	+	!		!
19. *T. lamutensis* Golovn.					+
20. *T. emimensis* Golovn.					+
21. *Trochodendrocarpus arcticus* (Heer) Kryshtofovich	+		+		+
22. *Platanus raynoldsii* Newberry	!	+	+	!	!
23. *Corylus beringiana* (Krysht.) Golovn.	!	!	!	!	+
24. *Cissites volobuevae* Herman		+	+		+
25. *Zizyphus* (?) sp.		+			
26. *"Pterospermites" platanoides* Golovn.		+			+
27. *Liriophyllum aeternum* Golovn.			+		+
28. *Celastrinites septentrionalis* (Krysht.) Golovn.	+		+		!
29. *Cissus* sp.					+
30. *Sorbus* (?) sp.	+				
31. *Nyssa tschucotica* Golovn.		+			
32. *Rarytkinia* aff. *terechovae* Vassilevsk. et Golovn.	+		+		+
33. *Carpolithes biloculatus* Golovn.					+

Note: + – presence of a species; ! – predominance of a species.

semiaquatic herbs *Haemanthophyllum cordatum* Golovn. Among the dicotyledons, *Corylus beringiana* (Krysht.) Golovn. and a few species of *Trochodendroides* are abundant and predominant in most of the localities, *T. bidentata* Vassilevsk. et Golovn. prevailing among the latter. There are also frequent occurrences of *Platanus raynoldsii* Newb., *Cissites volobuevae* Herman, *"Pterospermites" platanoides* Golovn., *Celastrinites septentrionalis* (Krysht.) Golovn., and *Rarytkinia* aff. *terechovae* (Krysht.) Golovn. Finds of other angiosperms are not numerous.

The majority of the angiosperm species is characterized by high leaf polymorphism. In many cases, the extreme forms of the morphological series can be easily taken for different species and even genera. Noteworthy is a general large-leaved habitus of many angiosperms, particularly, *Trochodendroides bidentata* Vassilevsk. et Golovn., *Platanus raynoldsii* Newb., *Corylus beringiana* (Krysht.) Golovn., *"Pterospermites" platanoides* Golovn., *Liriophyllum aeternum* Golovn., and *Rarytkinia* aff. *terechovae* Vassilevsk. et Golovn.

In composition and main dominants, the Koryak taphoflora is very close to the early Rarytkin floral subassemblage found in the middle Rarytkin Subformation of the Rarytkin Ridge (Nesov and Golovneva, 1990). They differ only in a few species. The Koryak taphoflora has been correlated to some distant floras of the Far East and North America (Golovneva and Herman, 1992).

Opinions vary on the age of the Koryak floral assemblage. Vasilevskaya and Abramova (1974) considered it to be "... Maestrichtian or, most likely, early Maestrichtian (perhaps including the latest Campanian) ... ." Volobueva and Terekhova (1974) supposed that "... continental deposits of the upper Koryak Subformation could not be older than the Maestrichtian and, possibly, even do not include the lowermost Maestrichtian ..." and corresponded to marine deposits containing remains of *Inoceramus pilvoensis* Sok., *I. kusiroensis* Nag et Mat., and *Pachydiscus*. According to Vakhrameev and Akhmet'ev (1977), the Koryak floral assemblage is of Danian age. Recently, Golovneva suggested the late Maestrichtian age of the Koryak taphoflora (Nesov and Golovneva, 1990). I support this opinion, and the new data provides better arguments.

The youngest fossil mollusks found in the middle Koryak Subformation underlying the flora-bearing beds (not at the top of the subformation) are *Inoceramus pilvoensis* Sok., *Inoceramus* sp., and *Patagiosites alaskensis* Jones, of early Maestrichtian age (in Pokhialainen's opinion).

The Koryak Formation is overlain with or without a little stratigraphic disconformity by the Amaam Formation near the Amaam Lagoon and by the Chukotka Formation at the Ugol'naya Bay (Table 2). Some researchers (Dundo *et al.*, 1974) do not distinguish the Amaam Formation as an independent unit and suggest that the Chukotka Formation is distributed in the Amaam Lagoon area as well. The lower parts of these formations contain remains of foraminifers, mollusks, crinoids, spores, and pollen. Volobueva, who united the lower Amaam and lower Chukotka subformations into the Il'nai Horizon and studied the paleontological remains, suggested the Danian-Paleocene age of this horizon (Volobueva and Krasnyi, 1979). The paleontological data presented by Dundo (Dundo *et al.*, 1974), however, dates its lower part as the Maestrichtian-Danian. In the basal beds of the Chukotka Formation on the Al'katvaam River (Fig. 1), Dundo found *Parallelodon* sp. and fragments of an inoceram prismatic layer, indicative of the Late Cretaceous age of these deposits. Therefore, if the correlation between the basal parts of the Amaam and Chukotka formations is correct and the molluscan remains have not been redeposited (there are no doubts about this as yet), the Koryak flora may not be younger than the end of the Maestrichtian. It existed, most likely, during the middle-late Maestrichtian, using its threefold division (Nesov and Golovneva, 1990). This is in accordance with the occurrence of a marine vertebrate assemblage of supposedly late Maestrichtian age near the top of the Koryak Formation at the Ugol'naya Bay (Nesov and Golovneva, 1990).

The paleobotanical evidence affords more accurate dating of the Koryak taphoflora. As noted earlier, it is most close to the Rarytkin floral assemblage (Table 2) from the upper half of the Rarytkin Formation in the northern Koryak Highland (Nesov and Golovneva, 1990). This assemblage is far richer in composition (63 species) and can be subdivided into two subassemblages. The Koryak taphoflora corresponds to the earlier one, in spite of the presence of some species of the later subassemblage (*Trochodendroides bidentata* Vassilevsk. et Golovn., *T. lamutensis* Golovn., and *"Pterospermites" platanoides* Golovn). The lower half of the Rarytkin Formation, exposed at the Gornaya River (the left bank of the lower course of the Velikaya River), contains the Gornaya floral assemblage that immediately preceeds the Rarytkin assemblage and is linked with it by gradual transition (Nesov and Golovneva, 1990). There are, unfortunately, no interbeds of marine origin in the thick, Rarytkin flora-bearing strata to check dating the mentioned taphofloras. However, the Gornaya assemblage is close in composition to the reliably dated Kakanaut taphoflora from the basin of the Kakanaut River, near Lake Pekul'neiskoe (Nesov and Golovneva, 1990) (Table 2). In this section, a flora-bearing member, 500 m thick, overlies tuffaceous siltstones containing remains of *Pachydiscus subcompressus obsoletus* Mat., *Inoceramus pilvoensis* Sok., *I. kusiroensis* Nag. et Mat., *I.* ex gr. *shikotanensis* Nag. et Mat., and other mollusks, indicative of the second half of the Maestrichtian (Volobueva and Krasnyi, 1979). Remains of dinosaurs of the Hadrosauridae and Troodontidae families were found at the uppermost flora-bearing level on the Kakanaut River. These remains indicate that the Kakanaut flora is not younger than the end of the Cretaceous. Nesov and Golovneva (1990) studied the Kakanaut flora and reptiles and sug-

Table 2. Correlation of the Maestrichtian sections in the northeastern Koryak Highland (the main paleontological finds are given in parentheses)

Stage	Zone, bed (Pokhialainen, 1988; Pokhialainen and Byalobzheskii, 1984);	The Khatyrka-Opuka interfluve (Pokhialainen and Byalobzheskii, 1984);	The Emima-Il'naivaam interfluve (Volobueva and Krasnyi, 1979; Volobueva and Terekhova, 1974) with supplements	The Ugol'naya Bay (Volobueva and Terekhova, 1974; Dundo *et al.*, 1974)	The Rarytkin Ridge (Nesov and Golovneva, 1990)	Lake Pekul'neiskoe (Volobueva and Krasnyi, 1979; Nesov and Golovneva, 1979)
Danian-Eocene		Bystraya Formation (*Korjakia kociubinskii* Poch., *Pachydiscus* aff. *gollevillensis* Orb.)	Amaam Formation (*Glycymeris amaamensis* Volobueva)	Chukotka Formation (*Parallelodon* sp., *Inoceramus*)	Rarytkin Formation: Upper Subformation (Rarytkin flora)	Danian-Paleocene deposits (*Glycymeris* sp.)
Maestrichtian	*Korjakia kociubinskii*					
	Inoceramus (*Spiridoceramus*) *kusiroensis*	(*Inoceramus kusiroensis* Nag., *I.* ex gr. *balticus* Boehm.) Kul'kai Formation	Koryak Formation: Upper Subformation (Koryak flora)	Koryak Formation: "Podsopochnyi" coal bed (rare plant remains)	Rarytkin Formation: Middle Subformation (Rarytkin flora) (Gornaya flora)	Maestrichtian deposits: (dinosaurs of the *Troodontidae*, *Hadrosauridae*) (Kakanaut flora) (*Inoceramus kusiroensis* Nag., *I. pilvoensis* Sok.)
			Middle Subformation (*Inoceramus pilvoensis* Sok.)	Marine deposits (*Inoceramus* ex gr. *balticus* Boehm., *Canadoceras* sp.)	Lower Subformation (Gornaya flora)	
Campanian	*Inoceramus* (*Cataceramus*) *pilvoensis*					
	I. (*Mytiloides*) *naumanni schmidti*		Lower Subformation			?

gested their middle Maestrichtian age. As the Koryak taphoflora, which is of the same type as the early Rarytkin flora, is younger than the Kakanaut flora, which is of the same type as the Gornaya flora, it must be of late Maestrichtian age (Table 2).

The fact that two Cretaceous floras of different ages, the Gornaya and Koryak, exist at the level of the late Maestrichtian *Inoceramus kusiroensis* Zone and above it must cast some doubt upon their datings because of an insignificant duration of this interval. In this respect, two essential circumstances should be noted.

First, the *Inoceramus kusiroensis* Zone cannot, apparently, be considered to be the Uppermost Cretaceous deposits in the eastern Koryak Highland. According to Pokhialainen and Byalobzheskii (1984) and Pokhialainen (1988), the Kul'kai Formation is overlain by the Bystraya Formation in the Khatyrka-Opuka interfluve (Table 2). The upper part of the Kul'kai Formation contains molluscan remains from the *Inoceramus* ex gr. *balticus-I. kusiroensis* Zone; the lowest deposits of the Bystraya Formation contain fossil mollusks including *Pachydiscus* aff. *gollevillensis* Orb. and *Korjakia kociubinskii* Poch. "... The former show the Maestrichtian age of the enclosing rocks; the latter ... indicate that the *I. kusiroensis* group was not of the youngest Cretaceous inoceramids in the Far East" (Pokhialainen and Byalobzheskii, 1984, p. 67). Having studied the deposits analogous to the Bystraya Formation in other regions of the Koryak Highlands, these authors concluded that "... sediments with *Korjakia* and *Pachydiscus* often transgressively overlie the beds with *I. kusiroensis* and indicate wide distribution of the uppermost Maestrichtian marine facies in the studied region" (Pokhialainen and Byalobzheskii, 1984, p. 73). It should not be ruled out that a part of the flora-bearing deposits of the upper Koryak Subformation and/or the basal parts of the Amaam and Chukotka formations correspond in age to these uppermost Maestrichtian beds (Table 2).

Secondly, the existence of two different floras during the middle-late Maestrichtian can be explained by significant climatic fluctuations, which could not but affect the terrestrial vegetation. Thus, the composition of the Maestrichtian molluscan fauna from the southern Sakhalin suggests a considerable cooling approximately at the middle-late Maestrichtian boundary (Zakharov *et al.*, 1984). More thermophilic Kakanaut-Gornaya flora was likely to exist in the middle Maestrichtian (Nesov and Golovneva, 1990). The appearance of the Koryak (and Rarytkin) flora was apparently related to the late Maestrichtian cooling.

REFERENCES

Dundo, O.P., Vasilevskaya, N.D., Korotkevich, V.D., and Vasilenko, L.V., New Data on the Age and Stratigraphic Range of the Chukotka Formation in the Eastern Koryak-Anadyr' Region, *Stratigrafiya i Litologiya Melovykh, Paleogenovykh i Neogenovykh Otlozhenii Koryaksko-Anadyrskoi Oblasti*, NIIGA, Leningrad, 1974, pp. 58 - 64.

Golovneva, L.B. and Herman, A.B., New Data on the Age and Composition of the Flora from the Koryak Formation, *Botan. Zh.*, 1992, vol. 77, no. 7, pp. 60 - 71.

Nesov, L.A. and Golovneva, L.B., Evolution of Flora and Vertebrate Fauna and Climatic Fluctuations during the Late Senonian in the Northeastern Koryak Highland, *Kontinental'nyi Mel SSSR*, DVO Akad. Nauk SSSR, Vladivostok, 1990, pp. 191 - 212.

Pokhialainen, V.P., Inoceramids and the Cretaceous Biota from the North Pacific Ocean, *Doctorate (Geol.-Min.) Dissertation*, Institute of Geology and Geophysics, Siberian Branch, USSR Academy of Sciences, Novosibirsk, 1988, p. 33.

Pokhialainen, V.P. and Byalobzheskii, S.G., New Data on the Late Maestrichtian Biostratigraphy in the Eastern Koryak Upland, *Paleontologiya i Biostratigrafiya Koryakskogo Nagor'ya*, SVKNII DVNTs Akad. Nauk SSSR, Magadan, 1984, pp. 64 - 74.

Vakhrameev, V.A. and Akhmet'ev, M.A., Higher Plants according to the Leaf Study, *Razvitie Flor na Granitse Mezozoya i Kainozoya*, Nauka, Moscow, 1977, pp. 39 - 65.

Vasilevskaya, N.D. and Abramova, L.N., The Late Cretaceous Floral Assemblages from the Koryak-Anadyr' Region, *Stratigrafiya i Litologiya Melovykh, Paleogenovykh i Neogenovykh Otlozhenii Koryaksko-Anadyrskoi Oblasti*, NIIGA, Leningrad, 1974, pp. 31 - 37.

Volobueva, V.I. and Krasnyi, L.L., *Maastrikht-Neogenovye Otlozheniya Vostochnoi Chasti Koryakskogo Nagor'ya* (The Maestrichtian-Neogene Deposits of the Eastern Koryak Highland), Nauka, Moscow, 1979.

Volobueva, V.I. and Terekhova, G.P., On the Cretaceous-Paleogene Boundary Beds in the Eastern Koryak-Anadyr' Region, *Stratigrafiya i Litologiya Melovykh, Paleogenovykh i Neogenovykh Otlozhenii Koryaksko-Anadyrskoi Oblasti*, NIIGA, Leningrad, 1974, pp. 53 - 57.

Zakharov, Yu.D., Grabovskaya, V.S., and Kalishevich, T.G., The Late Cretaceous Succession of Marine Associations in Southern Sakhalin and the Climatic Characteristics of the Northwest Pacific, *Sistematika i Evolyutsiya Bespozvonochnykh Dal'nego Vostoka*, DVNTs Akad. Nauk SSSR, Vladivostok, 1984, pp. 41 - 90.

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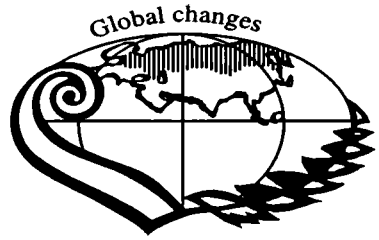

Evolution of the biosphere

Floral Stages in the Paleogene of Kamchatka: Sequence and Active Factors

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Abstract – The sequence of typical floras in the Paleocene and Eocene of Kamchatka is represented by their brief characteristics and an analysis of floral evolution under the impact of climate and other factors. The changes in floras show the influence of high ranking temperature fluctuations and the destabilizing effect of contrasts in climate, which grew since the middle of the Eocene; floral changes also record a higher tectonic and volcanic activity, which intensified the processes of structural and taxonomic renovation of the floras. As a result, by the end of the Oligocene, this process reached the stage when the floras ultimately lost their connection with the North Pacific ancient flora of the Paleogene.

The Cenozoic rocks of Kamchatka abound with the fossilized remains of ancient biota, thus making this region interesting for historical and evolutionary research. The possibilities here are unique for the study of the history of the Paleogene flora in the North Pacific, which requires the establishment of a real sequence of local floras in the combined section of the Paleogene, which is not always easy in this large and geologically complicated region. During the last decade, the geologists, paleontologists, and paleoflorists from the Geological Institute of the Russian Academy of Sciences and Moscow State University studied the Paleogene stratotypes in the type localities of Western Kamchatka. Their research showed that a series of errors in the datings and correlation of subdivisions of different rank occurred because of a regional stratigraphic hiatus in the Eocene section, not acknowledged in earlier research. The hiatus proved to be the result of a phase of a powerful tectogenesis (Gladenkov *et al.*, 1985; Shantser *et al.*, 1985). The concept of a continuous Paleogene section had a negative effect on phytostratigraphic and floragenetic interpretations, because it erroneously correlated the upper flora-bearing part of the Chemurnaut Group (Tkapravayam Formation) of the Penzhina region with the Snatol Horizon on the regional scale (Resheniya ..., 1982). This assumption essentially distorted the historical sequence of events. We have already mentioned this error from the point of view of the global climatostratigraphy of the Paleogene (Chelebaeva and Bratseva, 1985). Later, Paleocene (perhaps partly early Eocene) age of the Tkapravayam flora was confirmed by the finds of the late Danian foraminiferal assemblage at the base of the Chemurnaut Group (Serova, 1989). The study of flora from the middle Eocene Snatol Formation in the Tigil' locality revealed its principal difference from the Tkapravayam flora (Gladenkov *et al.*, 1991). The combined geological and biostratigraphical studies conducted by the Geological Institute in the Tigil' type locality, in the Paleogene stratotypes of the Palana and Penzhina localities, and on Il'pinskii Peninsula produced data for more accurate definition and resolution of the existing regional stratigraphic and correlative scales (Gladenkov *et al.*, 1989), including the phytostratigraphic scheme.

In the Paleogene of western Kamchatka, we define a sequence of eight floral assemblages (or types of floras) reflecting different stages in the paleoclimatic and paleogeographic environmental changes. Most of them are traced laterally and can be regarded as floral horizons. This sequence, however, has two large lacunae due to a lack of data, i.e., the interval of the low-middle Eocene stratigraphic hiatus and an interval of the Oligocene marine transgression. A description of these stages is given below.

(1) *The early Ulevenei stage.*[1] The type flora of this stage originates from the Cape Zubchatyi Formation established on the Utkholokskii Peninsula (Chelebaeva and Shantser, 1988). This flora contains plant assemblages of wide valleys and coastal flatlands framing

[1] Named after the Ulevenei series of the Tigil' locality, which includes the Cape Zubchatyi Formation and the Khulgun and Sosopkhan formations.

a large, brackish-water lagoon isolated from the sea, according to the reconstructions based on taphonomic, taxonomic, and lithological-facies data. The angiosperms are dominant in the flora; among them, wide-leaved plants with dentate blades prevail, i.e., the leaves are of medium size (5 - 10 cm), seldom up to 20 cm. Among the dominant plants are *Cocculus*, *Trochodendroides*, *Platanus*, *Populus*, *Cordia* ("*Grewiopsis*"), and representatives of conifers (*Taxodium*, *Metasequoia*). Single imprints of *Ulmus*, *Alangium*, and *Aesculus* were found. The habitus aspect of the flora and its composition imply a rather humid and warm-temperate climate.

The flora from the stratotype of the Khulgun Formation on the Kvachina River has similar characteristics, including dominant angiosperms with dentate, medium-size leaves. However, *Platanus* and *Populus* are not so frequent, whereas *Tetracentron* and the large-leaved *Aesculus*, which are particularly typical of the next stage, are more numerous. These features may suggest the intermediate position of flora in the Ulevenei Group section and allow considering it as a late display of the early Ulevenei flora. The appearance in the upper part of the stratotype of the Khulgun Formation of interbeds with brackish sea fauna most likely implies imperfect isolation of the Ulevenei basin from the sea.

(2) *The late Ulevenei stage.* The type flora originates from the Sosopkhan Formation, which crops out in the middle course of the Snatol River (Tigil' locality), and whose deposits were earlier referred to the Khulgun and Napan formations (Resheniya ... , 1982). The status of this unit should be modified, in light of the absence of both geological and floral reasons for its subdivision (Chelebaeva and Shantser, 1988). Tectonic dislocations preclude a reliable definition of the relationship among the large fragments of the section and the species distribution in it. On the whole, the formation is characterized by an alternation of coastal-marine and subcontinental facies, with a predominance of the latter (coastal flatlands, delta flood plains, etc.). The most striking peculiarity of the Sosopkhan flora is the large size of the leaves of many angiosperms and conifers; e.g., often the leaves are 20 - 30 cm long and even longer, and a form of *Metasequoia* has leaves double their normal size. The flora still retains most of the early Ulevenei genera; however, they are much more diverse. Dominant are *Metasequoia*, *Tetracentron*, *Cocculus*, *Alnus*, *Aesculus*, and *Acer arcticum*; more scarce are *Platanus*, *Corylus*, *Populus*, *Myrica*, and *Cordia*. There are single finds of entire leaves of *Magnolia* and *Lasianthus* and of other thermophilic dicotyledons and imprints of the fern *Asplenium nilssonioides*. The presence of the latter and of the large-leaved *Myrica beringiana*, common to thermophilic floras of the Chemurnaut Group in the Penzhina locality, implies that the late Ulevenei Horizon can be correlated with a certain part of the Chemurnaut section, preferably with the interval of the Getkilnin and Kamchik formations.

The late Ulevenei flora obviously enjoyed a more humid and warmer climate than the early Ulevenei flora, though the climate can hardly be called subtropic because of an insignificant amount of evergreen forms with entire leaves.

(3) *The Chemurnaut stage.* The type flora originates from the deposits of the upper part of the Chemurnaut Group stratotype in the Penzhina locality. The member has a distinctly regressive character and is subdivided into the Getkilnin (marine facies), Kamchik, and Tkapravayam formations (coastal-marine and continental facies). The latter two include deposits of beaches, coastal lowlands, and parts of large deltas. The upper part of the Tkapravayam Formation is composed of exclusively river alluvial deposits. The lower boundary of the Formation is largely conventional, and most of the researchers interpret it at random. This is why one can speak only about the undivided Chemurnaut flora in the section prior to the final study of all material. Large leaves and gigantism are typical of Chemurnaut flora to a still greater degree; i.e., the spectrum of thermophilic plants in it is more diverse and includes a larger number of forms with entire leaves than the late Ulevenei flora, with their role increasing up-section. A large part of temperate genera from the Sosopkhan flora is also preserved, but their role reduces in the same direction except, perhaps, that of *Platanus* and *Cocculus*. Among the dominant plants are *Cocculus*, *Myrica beringiana*, *Luvunga*, *Cordia*, *Platanus*, *Saurauia*, and *Dombeya*. South of the stratotype, in the upper part of the section, the leaves of the palm *Sabal* were found (Budantsev, 1979). The presence of the latter, along with several genera of the family Lequminosites (*Saurauia*, *Dombeya*, *Myrtophyllum*, and others), implies a humid, subtropical climate.

On the basis of the general climatostratigraphic concepts and comparisons with the floras of other regions, the age of the analyzed floras was determined as the Danian (early Ulevenei), late Paleocene (late Ulevenei), and late Paleocene-beginning of the early Eocene (Chemurnaut). The gradual change in plant composition in that interval relates most likely to the initially high general level of the thermal regime and the slow growth of paleotemperatures, with a simultaneous increase in the humidity, which thus maintained a mild climate without sharp contrasts. The warming was not accompanied by dessication, as in the Late Cretaceous in northwestern Kamchatka (Herman *et al.*, 1991). The cause of this could have been the relief reorganization during the Laramian tectogenesis, when mountain barriers rose on the side of the ancient Okhotsk land, thus reducing the influence of the continent. This was further supported by a more intensive clastic material supply from the area, which is recorded in the Chemurnaut Group sediments.

The post-Chemurnaut tectogenesis ended in a general uplift and turned the larger part of the region into an intensively eroded area. Sediment deposition seems to have continued only in a few, mostly peripheral, basins. This process might be responsible for the absence in the Kamchatka region of the buried flora of "paratropical" type, dated as the early Eocene

(the Lower Ravenian flora of the Kushtaka Formation), whereas in Alaska, this flora follows the Paleocene flora, which is similar to the Chemurnaut flora.

Another cycle of intensive deposition in Western Kamchatka started in the middle Eocene, when in a series of large marine basins fringing the Cretaceous and Paleocene rock, prominences appeared. In the lower part of this depositional cycle, composed of the Snatol and Kovachina horizons, there are several stages showing later changes in paleofloras. Their description is given in a special paper (Gladenkov *et al.*, 1991) and, therefore, we shall only name them: the Ust'-Anadyr', lower Snatol, upper Snatol, and Kovachina phytohorizons. The Snatol-Kovachina floras inherited the Paleocene features by preserving the major warm-temperate components, particularly in the lower two horizons: *Taxodium*, *Metasequoia*, *Cocculus*, *Trochodendroides*, *Platanus*, *Acer arcticum*, *Aesculus*, and *Cordia*. The Ust'-Anadyr' flora also contains thermophilic relics of the ancient optimum flora: *Haemanthophyllum* and *Calkinsia*, which disappear in higher horizons. The lower Snatol flora implies a rather unfavorable climate and is considered to prefer a cold environment, because the sediments contain glendonites, i.e., concretions whose formation and preservation require low temperatures. Their abundance and frequency of occurrence increase in the overlying part of the section, despite the growth of the abundance and diversity of thermophilic plants in the lower Snatol-Kovachina beds. This suggest the appearance of cold seasons and, consequently, the formation of a new type of contrasting climate with cold winters, most likely reflecting a global process. The upper Snatol and the Kovachina floras show traces of essential transformations in vegetation influenced, first, by the onset of a warming and, secondly, by the appearance of a high-mountain volcanic relief, which in a contrasting climate, initiated vertical zoning. In fossil assemblages, these events produced a larger share of conifers (Pinaceae, in particular) and their high taxonomic diversity. A typical feature of the Late Snatol-Kovachina flora is the presence in it of abundant *Sciadopitys*, *Cunninghamia*, *Sequoia*, *Cryptomeria*, the genera of Lauraceae, and several other forms indicative of an almost subtropical climate.

The determination of climatic conditions according to the amount of forms related to the corresponding modern plant assemblages is largely arbitrary. There is an obvious consistency in the difference between the floras that were "close to subtropical" during the time of the frostless and uncontrasting climate (the Chemurnaut stage) and the recent-type floras of the contrasting climate (late Snatol-Kovachina flora); this consistency permits no age correlation of deposits with these floras (the Tkapravayam Formation and the Snatol deposits).

The onset and intensification of a contrasting climate with frosty seasons conditioned the later development of floras of the Northern Pacific (as also, probably, on a global scale); moreover, this kind of climate increasingly affected the geographical differentiation of floras and intensified the influence of other paleogeographical and geological factors. As yet, there are no sufficiently representative findings of Oligocene plants in Kamchatka. Single finds of *Thuja* and *Larix* in marine sediment stratotypes supply the only information.

The flora of the Krutogorova River locality (Chelebaeva, 1988), referred to the Oligocene-lower Miocene boundary beds, is of mesophilic warm-temperate aspect, having little in common with the Eocene floras. With reference to the late Eocene floras, it should be noted that the age of the Snatol and Kovachina horizons, according to malacofaunas and foraminifers, was confined to the middle-late Eocene (Gladenkov *et al.*, 1991). It seems the Kovachina phytohorizon, however, does not completely cover the late Eocene. Many paleoflorists consider the warm-temperate mesophilic Irgirnin flora from the formation of the same name in Podkagernaya Bay as the latest flora of this time period. We defined the Irgirnin stage as just above the Kovachina (Gladenkov *et al.*, 1989).

The composition and structure of the Irgirnin type flora has many common features with the late Snatol-Kovachina floras; e.g., it has abundant and diverse conifers, particularly *Sequoia*, with the genera of the ancient temperate flora, such as *Cocculus*, *Platanus*, and *Cordia*, still playing significant role. A characteristic feature of the Irgirnin flora is the presence of genera of later time periods, the Oligocene and Neogene (*Fagus*, *Castanea*, *Alnus*, *Juglans*, *Comptonia*, and others), and therefore, it can be considered as post-Kovachina (or late Kovachina?). We should be careful in doing so because of the taphonomic peculiarity of the Kovachina-type flora (burial in entirely marine facies) with incomplete composition.

The taphonomic factor is especially important while studying the late Eocene floras of Kamchatka because of the higher regional volcanic activity in late Snatol and Kovachina time. Single or repeated eruptions have covered various areas with volcanogenic-sedimentary beds and with igneous rocks, thus burying the plant assemblages differing from the climax floras, including pioneer ones. In certain cases, these plant assemblages can be assumed as "impoverished" assemblages of a cooler period; they are frequently observed in the volcanites of the Kinkil' Formation and in its analogs, whose stratigraphic position in the interval of the Snatol and Kovachina horizons is not always clear.

Therefore, the Paleogene flora of Kamchatka reveals two distinct periods of changes: the Paleocene-early Eocene and the middle-late Eocene. Climate was the principal factor in the changes in vegetation in the first period, and climate and volcanism were most important in the second period, with the latter affecting flora both indirectly (relief) and directly (partial destruction).

The slow rise of paleotemperatures in the first period increased the humidity, apparently because of a barrier that appeared in the relief of continental Asia. In

the second period, the seasonal contrasts in climate grew against the background of long-term variations of the temperature regime.

The long-term megathermal regime in the first period was favorable for the formation of taxonomically rich large-leaved thermophilic flora in the vast territories of the continental framing of the Northern Pacific and on the adjacent high-latitude areas. The uniformity of this flora is often an obstacle to its stratification, which requires abundant fossil material.

In the second period, the tectonic instability in post-Chemurnaut time, the appearance and intensification of seasonal contrasts of climate, the oscillations of paleotemperatures, and, finally, the activization of volcanism enhanced the processes of structural reorganization and renovation in the systematic composition of flora in Kamchatka. By the end of the Oligocene, this flora lost all links with the ancient Paleogene floras, thus launching the desintegration of a single North Pacific flora.

REFERENCES

Budantsev, L.Yu., The Find of the Palm *Sabal* in the Eocene Deposits of Northwestern Kamchatka, *Botan. Zh.*, 1979, vol. 64, no. 12, pp. 1777 - 1783.

Chelebaeva, A.I., The Late Oligocene-Early Miocene Flora of Kamchatka, *Ocherki po Geologii Kamchatki i Koryakskogo Nagor'ya*, Nauka, Moscow, 1988, pp. 25 - 68.

Chelebaeva, A.I. and Brattseva, G.M., On Climatostratigraphy of Paleogene from Paleofloras (North Pacific and Adjacent Regions), *Korrelyatsiya Kainozoiskikh Tolshch Dal'nego Vostoka*, Nauka, Moscow, 1985, pp. 157 - 209.

Chelebaeva, A.I. and Shantser, A.E., New Data on Early Paleogene of Western Kamchatka, *Litologiya i Stratigrafiya Mezozoya i Kainozoya Vostochnykh Raionov SSSR*, Nauka, Moscow, 1988, pp. 135 - 148.

Herman, A.B. and Lebedev, E.L., *Stratigrafiya i Flora Melovykh Otlozhenii Severo-Zapadnoi Kamchatki* (Stratigraphy and Flora of Cretaceous Deposits in Northwestern Kamchatka), Nauka, Moscow, 1991.

Gladenkov, Yu.B., Brattseva, G.M., Sinel'nikova, V.N., *et al.*, The Stratigraphic Scheme of Paleogene of the Far East, *Kainozoi Dal'nego Vostoka*, DVNTs Akad. Nauk SSSR, Vladivostok, 1989, pp. 161 - 168.

Gladenkov, Yu.B., Sinel'nikova, V.N., Chelebaeva, A.I., and Shantser, A.E., Problems of Paleogene of the Tigil' Type Locality in Western Kamchatka, *Korrelyatsiya Kainozoiskikh Tolshch Dal'nego Vostoka*, Pt. 1, Nauka, Moscow, 1985, pp. 45 - 55.

Gladenkov Yu.B., Sinel'nikova, V.N., Shantser, A.E., and Chelebaeva, A.I., *Eotsen Zapadnoi Kamchatki* (The Eocene of Western Kamchatka), Nauka, Moscow, 1991.

Resheniya 2-go Mezhvedomstvennogo Stratigraficheskogo Soveshchaniya po Melu, Paleogenu i Neogenu Koryakskogo Nagor'ya, Kamchatki, Komandorskikh Ostrovov i Sakhalina (Resolutions of the 2nd Interdepartmental Meeting on Stratigraphy of the Cretaceous, Paleogene, and Neogene of the Koryak Highland, Kamchatka, Commandor Is., and Sakhalin), PGO "Kamchatgeologiya," 1974, Petropavlovsk-Kamchatskii, 1982.

Serova, M.Ya., Foraminifers of Marine Deposits of the Danian Stage in Northwestern Kamchatka, *Voprosy Mikropaleontol.*, 1989, no. 30, pp. 66 - 78.

Shantser, A.E., Chelebaeva, A.I., and Kraevaya, T.S., A Tectonic Episode in the Eocene of Northwestern Kamchatka and Its Stratigraphic Implication, *Korrelyatsiya Kainozoiskikh Tolshch Dal'nego Vostoka*, Pt. 1, Nauka, Moscow, 1985, pp. 19 - 44.

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Evolution of the biosphere

Meridional Distribution of Pleistocene Ecosystems in Asia: Basic Problems

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Abstract – The basic problems of meridional and altitudinal ("mountain") changes in the Pleistocene ecosystems of Asia are analyzed with reference to the cooling-warming (glacial-interglacial) cycle. Estimates are made of the duration of the cold and warm stages in various latitudinal and altitudinal zones, the migration of landscape zones, and the correlations between the chronostratigraphic and climatostratigraphic boundaries and between glacial-interglacial and pluvial and arid periods. The periglacial period and the "cold" and "warm" loesses are discussed in context with their spatial and age relationships. A model of meridional and altitudinal analysis of changes in the ecosystems is suggested.

In Quaternary time, as compared to the older stages of the Cenozoic, many new groups of ecosystems appeared with their own specific landscapes and sedimentation types. These changes in the ecosystems resulted from glacial, cryogenic, glacial-marine, pluvial, and other peculiar environments caused by general climatic changes, primarily, by an overall cooling. The new landscape-climatic zones of the Quaternary, i.e., glacial, periglacial, and pluvial, can be considered not only as paleogeographic phenomena of the Ice Age, but also as paleoecosystems with specific biotic and abiotic processes. In their development, the paleoecosystems are interrelated in both geological time and space as sensitive recorders of climatic changes. Therefore, the stratigraphy and paleogeography of the Quaternary actually form parts of the general paleoecosystem analysis. Although the problems of stratigraphy and paleogeography dominated in the Quaternary geology, this kind of analysis is as yet neglected. This paper suggests a variant of the general paleoecosystem analysis, based on the materials covering Asia, and discusses certain debatable problems of the stratigraphy and paleogeography of that period.

The stratigraphic analysis of the Quaternary is conducted on the same principles as that of other systems and is primarily concerned with the biostratigraphical basis for the identification of large Quaternary subdivisions, ranking them in series. In this case, the evolutionary principle of biostratigraphy is dominant. In more detailed stratigraphical subdivisions, however, the climatostratigraphic principle becomes the primary factor in Quaternary stratigraphy. Distinguishing the Quaternary as a new ecosystem of the Ice Age in the Earth's history has established the climatostratigraphical dominant as the systematic basis for its subdivision. In fact, since the Ice Age history was first propounded, all stratigraphical schemes were based on the determination of "cold" (glacial) and "warm" (interglacial) epochs or on their stage variants. The most promising in creating the glacial stratigraphy is the zone of conjunction between the extraglacial regions and the edges of ice sheets in Eastern Europe, western Siberia, and North America, because the glacial (moraine or fluvioglacial) beds and their analogs divided by interglacial (intermoraine) deposits are sufficiently distinguishable here. This is why these areas were and still are the key regions for the detailed stratigraphy of the Quaternary, in particular, of the Pleistocene.

It should be noted that the climatostratigraphic standards produced specifically for these middle latitude regions in the contact zone of the ice sheets were used as general stratigraphical schemes for the Quaternary. Their principles are used for other ecosystems of the landscape-climatic belt of the entire extraglacial zone, down to the tropics. This indiscriminate application of climatostratigraphic principles and standards, adopted for a concrete zone of the ice sheet boundary in the systematization of the more southern regions of the entire extraglacial zone, may involve errors of principal significance. Therefore, the general theoretical analysis of the history of changes in the Pleistocene natural environment during coolings and warmings in the different ecosystem landscape-climatic zones of the Northern Hemisphere is of extreme importance. In recent years, many publications were dedicated to the various aspects of this problem. I define it as "the problem

of meridional analysis and the correlation of paleogeographical events in the Pleistocene" (Pozdnii Kainozoi ... , 1989; Devyatkin, 1990, 1991; Devyatkin and Malaeva, 1991). Certain aspects of the problem cause not only debates but conflicting interpretations. Among them are the following:

(1) Estimation of the duration of cooling and warming stages in different latitudinal zones. The geochronological scale of the Quaternary for continental regions covers mainly the middle latitudes, i.e., the contact zones between the ice sheets and the adjacent regions of the extraglacial zone. "Geochronological standards," then, are applied to the more southern regions of the extraglacial zone, thus creating an illusion of a synchronous occurrence of the main climatic rhythms of the entire meridional profile. All theoretical climatological works show, nonetheless, that the durations of coolings and warmings in different latitudinal zones were different, which is well demonstrated in recent time and can be entirely referred to the paleoclimates of the past. The peaks of climatic cycles of coolings and warmings (minima and maxima of temperatures) actually show synchronism, but the duration of the stages themselves is different in different zones. Therefore, the time range of coolings and warmings should be different; i.e., the cooling periods are longer in high latitudes, whereas the warming stages are longer in lower latitudes. Certain "cold" and "warm" wedges are formed, whose climatostratigraphical boundaries are diachronous in relation to the chronostratigraphic boundaries and cut across them in a meridional direction. The theoretical model in Fig. 1 represents this statement well.

(2) Reconstruction of the paleoclimatic parameters of the cooling and warming (temperature, precipitation, and circulation) in different latitudinal zones reveals their difference. The intensity of the cooling or warming processes conditioned this difference, as did the corresponding reaction to them of the biotic and abiotic environment, reflected in the different types of landscape ecosystems, in their changes and the type of sediment deposition itself (glacial, periglacial, extraglacial, and pluvial), and in the peculiarities of their occurrence.

(3) The establishment of climatostratigraphic standards to distinguish the cold and warm stages in different landscape-climatic zones. This is an important issue, because palynological data, as the basic method in Pleistocene climatostratigraphy, may indicate that compositionally similar palynospectra correspond in different zones to diametrically opposite climatostratigraphic settings. A typical example: in the tundra sections, the palynospectrum with wood vegetation is interpreted as an indication of warming, whereas wood elements in the palynospectrum of the steppe zone are thought to reflect a cooling. Therefore, it appears that a set of typical palynological "standards" should be fixed for glacials and interglacials (i.e., coolings and warm-

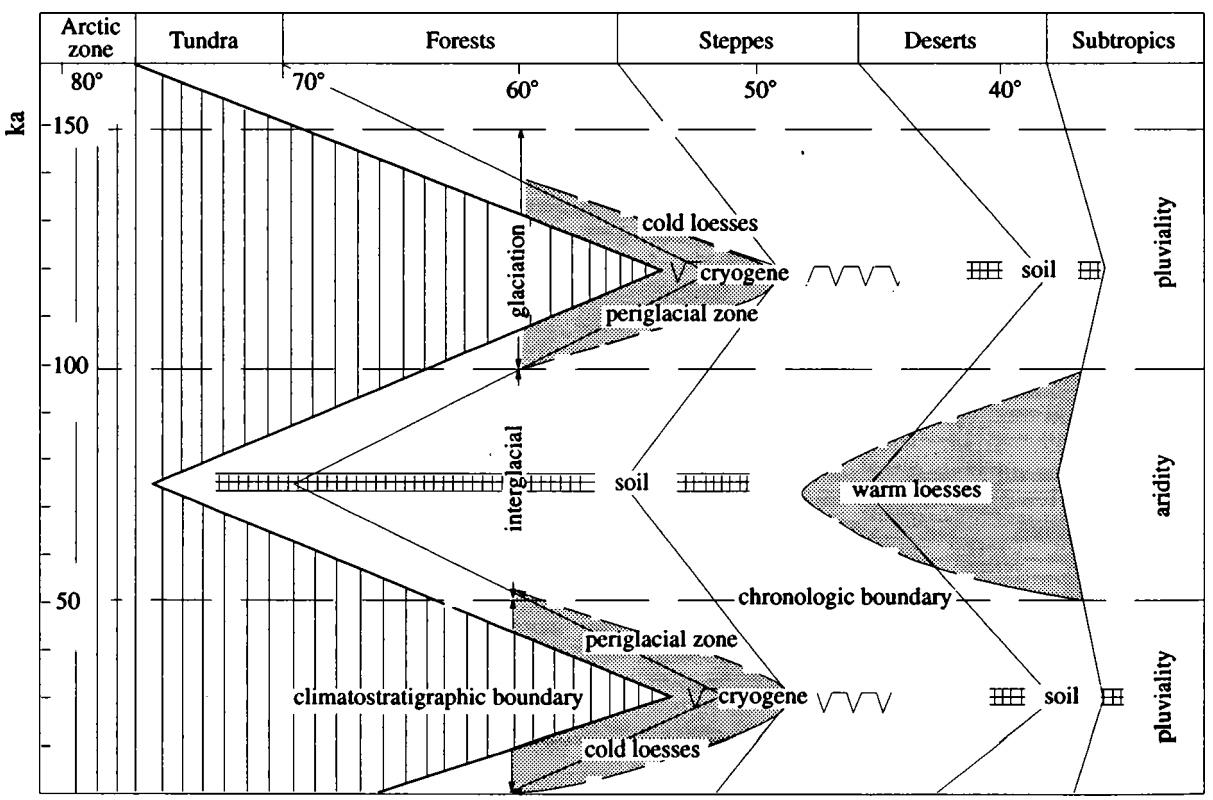

Fig. 1. *Theoretical model of the meridional analysis of events in the Pleistocene (the glacial-interglacial cycle): major processes and latitudinal migrations of landscape-climatic zones.*

ings) in different zones. The meridional profile can accurately trace lateral variations in the change of ecosystem vegetal associations for the same stratigraphic horizon.

(4) The meridional and vertical migrations of landscape-climatic zones that occurred as a result of climatic fluctuations, i.e., during cooling in the south and during warming in the north. The basic postulates of the migration-climatic conception were elaborated by S.L. Troitskii (1982), with reference to the northern regions of western Siberia. The plain character of this territory allows the researcher to analyze migration processes in their "purest" form. In the late Pleistocene, the natural zones of western Siberia, as estimated from palynological data by V.S. Volkova (1977), moved more than 1000 km to the south (Fig. 2). Conditions favorable for wood plants remained only on the latitude of the Kazakh hillocky area. The migration of landscape zones was forward-backward, which caused the repeated appearance of similar plant groups on the same latitudes. The scheme of migration of the vegetal zones of the extraglacial area in western Siberia during the carpologically determined Karginskii and Sartan times is reliably substantiated by radiological data (^{14}C) (Krivonogov, 1988).

In the mountain regions of eastern Siberia, in the Altai, Sayany, and Tuva mountain belts, and in Mongolia, the vertical migration of the landscape-climatic altitude bands is foremost during the cooling-warming cycle. This migration has the same direction as meridional migration; i.e., during cooling, the bands descend to lower hypsometric levels; during warming, they rise to higher ground. In these regions, the elements of meridional and altitudinal migrations form different combinations (Pozdnii Kainozoi..., 1989; Devyatkin and Malaeva, 1991). Figure 3 shows such a model composed from materials in Mongolia.

(5) Paleogeographic reconstructions of the periglacial zone of Pleistocene glaciations. This zone is important for the reconstruction of Pleistocene paleolandscapes and climates. There are two interpretations of this term. In the first version, the periglacial zone is placed immediately adjacent to the glacial zone and bears the direct impact of glaciation. Specific ecosystems are formed accordingly in this zone: the landscapes of cold tundra-steppe, a special type of fauna, a specific lithogenesis under conditions of permafrost processes, and loess formation. The second version allows for a wider interpretation of the term and implies the same paleogeographic elements but over a larger area and not directly connected with the periphery of ice sheets, for example, in eastern Siberia and Central Asia. The formation of the vast spaces of this hyperzone is attributed to the general effect of the Ice Age climate. It should be noted that the modern climate of Siberia and inner Asia is favorable to the formation and preservation of landscape ecosystems similar to those of the periglacial, i.e., the presence of regions with a sharply continental climate, when the summer-winter temperature gradient reaches 100°C, the winter temperatures drop to −50°C (Oimyakon and Ubsa Nur), and there is little precipitation (about 200 mm/y) and deep freezing of the ground, particularly of humid soils. This is true of the elevated levels of highlands and of the zones with a nival type of landscape. These ecosystems should be considered as modern formations and cannot be classified with paleoecosystems of the Ice Age. It might seem, therefore, that the "periglacial model," without proof or necessity, is often included in paleogeographic reconstructions of the entire northern half of continental Asia. Even the regions of Central Asia and China, where loess beds have been considered now for more than 50 years as belonging to the generation of "warm" loesses, are identified with periglacial formations, and the principles of "glacial" stratigraphy are applied to them as well.

I suggest as the most reasonable, distinguishing two generations of loess formations: (a) cold periglacial loesses directly adjoining the boundary of ice sheets and (b) warm loesses of the arid zone. They correspond to different paleoclimatic settings and can be entirely asynchronous.

A reassessment of the climatostratigraphic interpretation of the loess-soil series is now contemplated for the south of western Siberia. I.A. Volkov and V.S. Zykina (1982) distinguish three climatic regimes and emphasize that "fossil soils appeared in a climatic setting similar to the recent environment; their secondary transformations took place in a humid and cold climate, whereas loess or loesslike loam overlying the soil accumulated in very dry climate" (p. 26). V.S. Zykina (1990) points out that "the climatic rhythm shows a consecutive alternation in time of three different types of natural environments: dry and warm (formation of loess cover), similar to recent time (formation of fossil soils), and humid and cold (cryogenesis)" (p. 32).

The paleoclimatic reconstructions of the time of formation of the loess-soil series in Central Asia have also undergone corrections. M.M. Pakhomov (1983) gave a detailed analysis of the list of plants recovered from loess and soils at the two reference sections in the Tadzhik depression; he also indicated climatic parameters (temperature and precipitation) at the time of their formation. For example, in the Lakhuti section, as "indicated by climatic implications of the observed floral assemblages, the mean annual temperature at the time of formation of the V pedocomplex (PC) was 9 - 10°C. The loess overlying the V PC accumulated in warmer conditions, with a mean annual temperature of 12 - 13°C" (p. 971). M.M. Pakhomov had every reason to conclude that the accumulation of loesses "was an interglacial zonal indication of an area of dry subtropics" (p. 969), whereas the fossil soils were deposited in relatively cool and humid periods of glaciations.

Consequently, the new paleoclimatic data on the loess-soil series in southwestern Siberia and Central Asia refer them to the generation of "warm" loesses and correlate them with interglacial periods, whereas the soils correspond to glacial periods. This means that there is no direct correlation between loesses and soils

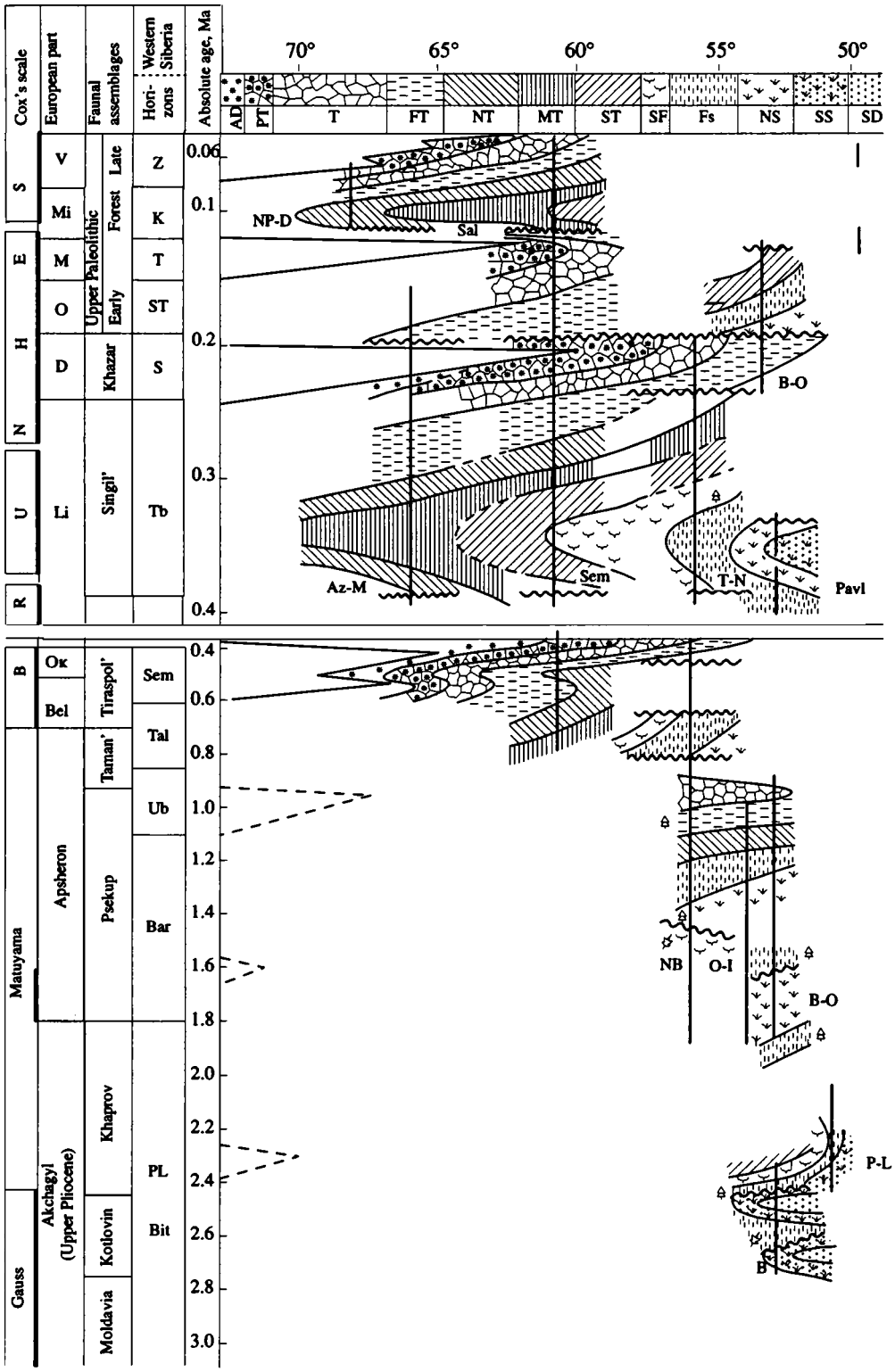

Fig. 2. *Migration of vegetational zones in the late Pliocene and Quaternary (after V.S. Volkova, 1977). Vegetational zones: AD – Arctic desert; PT – periglacial tundra; T – tundra; FT – forest-tundra; NT – northern taiga; MT – middle taiga; ST – southern taiga; SF – small-leaved forests; FS – forest-steppe; NS – northern steppe; SS – southern steppe; SD – semidesert.*

Horizons: V – Valdai; Mi – Mikulino; M – Moscow; O – Odintsovo; D – Dnieper; Li – Likhvin; Ok – Oka; Bel – Belovezh'e; Z – Zyryanka; K – Kazantsevo; T – Taz; ST – Samarovo-Taz; S – Samarovo; Tb – Tobol; Sem – Semeika.

Beds: Tal – Talagai; Ub – Uba; Bar – Barnaul; PL – Podpusk-Lebyazh'e; Bit – Biteke; Sections: Sem – Semeika-Samarovo; NP-D – Novyi Port-Dudinka; Sal – Salekhard; Az-M – Azov-Muzhi; T-N – Tomsk-Novosibirsk Ob' area; Pavl – Pavlodar Irtysh area; BP – Barnaul Ob' area; NB – Northern Baraba; B – Biteke; P-L – Podpusk-Lebyazh'e.

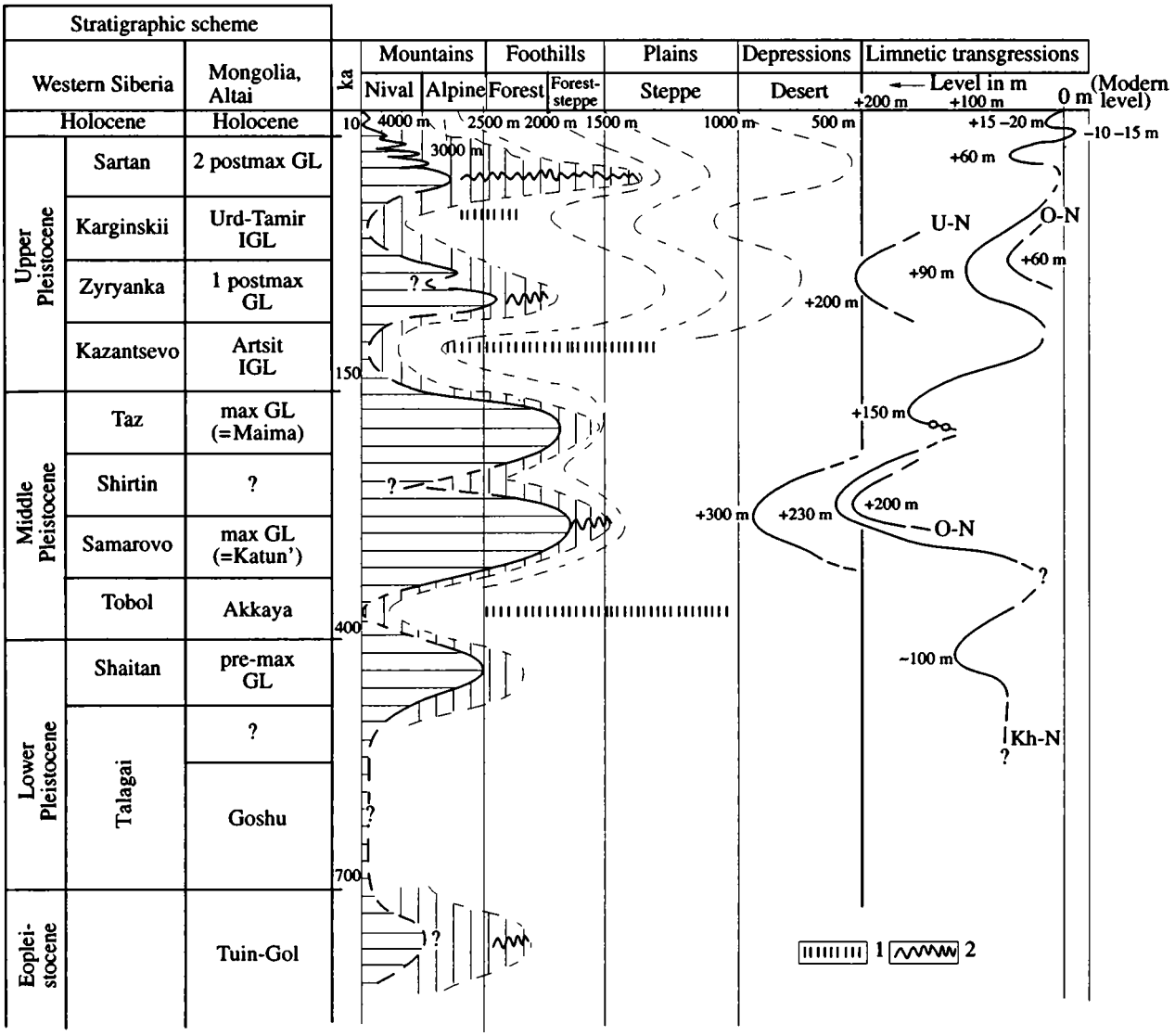

Fig. 3. *An analysis of ecosystems of the Eopleistocene and Pleistocene in the Altai–Sayany mountain area and in Mongolia: the vertical migrations of landscape-climatic zones, glacial and cryogenic processes, transgressions and regressions of pluvial basins (Kh-N – Khirtis-Nur; U-N – Ubsa-Nur; O-N – Orok-Nur).*

1 – soils; 2 – cryogenesis. Stratigraphic subdivisions: GL – glacials; IGL – interglacials.

of arid and periglacial zones, because they correspond to different paleoclimatic cycles and are asynchronous.

The periglacial problem is, no doubt, of great importance to the Pleistocene paleoclimatic reconstructions. In contrast to the "maximalistic" views on the periglacial zone in the papers on continental Asia, the research on North America, in my opinion, shows more objective approaches. The periglacial zone in the latter is clearly confined to the distribution of relict continuous and fragmentary permafrost at the edge of the Wisconsin glaciation (Prirodnye Usloviya ... , 1986). In the eastern part of the territory, this zone is not more than 200 km wide. A region of highland (Alpine) permafrost is particularly indicated, which, unlike the periglacial zone of glaciation, has no direct connection with the edge of the glacier.

(6) The problem of the correlation of glaciations and interglaciations (in high latitudes) with pluvial and arid periods (in low latitudes) is one of the debatable issues in the paleogeographical analysis of changes in the natural environment of the Pleistocene. I have written more than once on the subject, using the materials on inner Asia (Devyatkin, 1981, 1990; Devyatkin *et al.*, 1978; Pozdnii Kainozoi ... , 1989). It was recorded that coolings caused the pluvial band to move south into the arid zone, whereas during the warming periods, this band migrated to the north. In the arid zone of Asia, in the cooling periods, a single band of pluvial environments stretched from the Near East across the Caspian Sea, Aral, Balkhash, Alakul', and Sinkiang intermountain depressions, to western and eastern Mongolia (Murzaeva *et al.*, 1992). The band was characterized by more

humid climate and landscape conditions and by transgressions of drainless limnetic basins. The coolings in the zone created a specific ecosystem generally synchronous with the glacial and periglacial systems of the high latitudes. This observation, supported by the data on Asia, confirms the classical model elaborated by K. Butzer (1961) and R. Flint (1957), showing the synchronous occurrence of pluvial environments with glaciations and of arid conditions with interglaciations.

Actually, the problem is to correlate two climatostratigraphic models based on different climatic data, i.e., temperature for the high latitudes and humidity for the low latitudes.

In North America, the problem of pluvial and arid epochs has been discussed for more than a century. The publications dealing with the Late Quaternary natural environment in the USA clearly show that the concept of synchronous occurrence of cooling at high latitudes with pluvial environments in the arid zone, and vice versa, has solid grounds (Prirodnye Usloviya ... , 1986). It should be emphasized that these correlations are now based on representative radiological (^{14}C) data.

The problems discussed above are basic in the meridional analysis of Pleistocene ecosystems and, in particular, in the climatostratigraphy of glacial and extraglacial regions of Asia. At present, they are the most acute problems in the general theory of the dynamics of ecosystems in the geological past.

REFERENCES

Butzer, K.W., Climatic Change in Arid Regions since the Pleistocene, *Arid Zone Res.*, 1961, no. 17, pp. 17 - 52.

Devyatkin, E.V., *Kainozoi Vnutrennei Azii (Stratigrafiya, Geokhronologiya, Korrelyatsiya)* (The Cenozoic of Inner Asia), Nauka, Moscow, 1989.

Devyatkin, E.V., Problems of Meridional Correlation of Paleogeographical Events in the Pleistocene of Asia (Theoretical Aspects), *Tez. Dokl., VII Vsesoyuzn. Soveshch., Chetvertichnyi Period: Metody Issledovaniya, Stratigrafiya i Ekologiya*, Izd. Akad Nauk Estonii, Tallinn, 1990, vol. 1.

Devyatkin, E.V., Malaeva, E.M., Murzaeva, V.E., and Shelkoplyas, V.N., Pluvial Pleistocene Basins of the Great Lakes Depression in Western Mongolia, *Izv. Akad Nauk SSSR, Ser. Geograf.*, 1978, no. 5, pp. 37 - 51.

Flint, R.F., *Glacial and Pleistocene Geology*, Wiley and Sons, New York, Chapman and Hall, London, 1957.

Krivonogov, S.K., *Stratigrafiya i Paleogeografiya Nizhnego Priirtysh'ya v Epokhu Poslednego Oledeneniya* (Stratigraphy and Paleogeography of the Lower Irtysh Area in the Last Glaciation Epoch), Nauka, Novosibirsk, 1988.

Murzaeva, V.E., Sokol'skii, A.M., *et al.*, *Chetvertichnaya Geologiya Zarubezhnoi Azii* (Quaternary Geology on the Territory of Other Countries in Asia), Nauka, Moscow, 1991.

Pakhomov, M.M., New Data in Paleogeography of the Loess-Soil Series in Central Asia, *Dokl. Akad Nauk SSSR*, 1983, vol. 27, no. 4, pp. 967 - 972.

Pozdnii Kainozoi Mongolii (The Late Cenozoic of Mongolia), Nauka, Moscow, 1989.

Prirodnye Usloviya SShA v Pozdnechetvertichnoe Vremya. Pozdnii Pleistotsen (Natural Conditions on the USA Territory in the Late Quaternary), Gidrometeoizdat, Leningrad, 1986.

Troitskii, S.L., Basic Principles of the Migration-Climatic Conception, *Problemy Stratigrafii i Paleogeografii Pleistotsena Sibiri*, Nauka, Novosibirsk, 1982, pp. 6 - 10.

Volkov, I.A. and Zykina, V.S., Stratigraphy of Quaternary Deposits of the Loess Member in the Novosibirsk Ob' Area, *Problemy Stratigrafii i Paleogeografii Pleistotsena Sibiri*, Nauka, Novosibirsk, 1982, pp. 17 - 28.

Volkova, V.S., *Stratigrafiya i Istoriya Razvitiya Rastitel'nosti Zapadnoi Sibiri v Pozdnem Kainozoe* (Stratigraphy and History of Evolution of Vegetation in Western Siberia in the Late Cenozoic), Nauka, Moscow, 1977.

Zykina, V.S., The Natural Environment of Warmings in the Late Pleistocene from Paleopedological Data of Western Siberia, *Tez. Dokl., VII Vsesoyuzn. Soveshch., Chetvertichnyi Period: Metody Issledovaniya, Stratigrafiya i Ekologiya*, Izd. Akad Nauk Estonii, Tallinn, 1990, vol. 2.

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Structure and Lithologic-Mineralogic Features of the Basal Part of the Lower Riphean Deposits, Olenek Uplift

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Abstract – The geologic structure of the Lower Riphean Osor-Khayata Formation of the Olenek uplift, northern Siberia, is described, and detailed mineralogic characteristics of the glauconite widely occurring in its deposits are presented for the first time. It is shown that glauconite grains that have different morphology and dimensions but similar density properties are characterized by similar structural-crystallochemical specific features and are represented by Al-glauconite mica with a highly ordered structure. Various modes of generation and alternation for grains of globular and tabular shape are discussed.

The Lower Riphean section of the Olenek uplift, northern Siberia, is among the reference sections of this erathem in northern Eurasia. The structure of upper carbonate members of this section, delimited into the Kyutingda Formation, is at present indisputable, whereas two opinions exist on the subdivision of its lower, siliciclastic part. Some geologists, following the authors of the first stratigraphic schemes of the Olenek Riphean (Biterman and Gorshkova, 1962), recognize the Sygynakhtakh Formation alone (Yakshin, 1988), while others identify the lower, relatively fine-grained part of the silicoclastic sequence as an independent, Osor-Khayata Formation separated from the overlying deposits by an erosional surface and classify it either as belonging to the Lower Riphean (Kats and Frolova, 1986) or to the pre-Riphean part of the Proterozoic ("middle" Proterozoic) (Shpunt *et al.*, 1982).

This paper contains new evidence that confirms the independence of the Osor-Khayata Formation, describes its two most representative sections, and is the first to give the mineralogic characteristics of the globular and lamellar layer silicates of Al-glauconite compositions contained in the deposits of this formation.

The Osor-Khayata Formation is assigned to the Lower Riphean for two reasons: (1) its relatively close connection with the overlying Sygynakhtakh-Kyutingda succession, whose upper part (the Kyutingda Formation) includes a typical Lower Riphean set of stromatoliths, containing *Kussiella kussiensis* Kryl., *Iliella* cf., *Ilkotuicanica* Kryl., *Conophyton garganicus* Kor., and *Jacutophyton divulgatum Schap.* (Komar, 1966; Shpunt *et al.*, 1982) as interregional members; and (2) the correlation of Osor-Khayata, Sygynakhtakh, and Kyutingda deposits with paleontologically and isotopic geochronologically proven Lower Riphean deposits of the Anabar Massif (with the Mukun Group, Ust'-Il'ya and Kotuikan formations) and of the Siberian Riphean hypostratotype (with the Uchur Group), which is based on historical-geological and, for the upper carbonate horizons of these sequences, paleontological and isotopic geochronological data. The K-Ar dating of mineralogically unstudied glauconite from sandstones of Mt. Osor-Khayata, equal to 1435 Ma (Geokhronologiya Dokembriya ... , 1968) and initially thought to belong to the Sygynakhtakh Formation, was also regarded as evidence in favor of the Early Riphean age of the Osor-Khayata Formation. In this respect, it should be noted that some new isotopic datings of the Osor-Khayata Al-glauconite from our collection (including those from Mt. Osor-Khayata) obtained at the Institute of Precambrian Geology and Geochronology, RAS, have shown considerably lower, most likely rejuvenated values of the K-Ar and model Rb-Sr ages, which are beyond the limits of the Early Riphean, 1650 ± 50 to 1350 ± 20 Ma (Bibikova *et al.*, 1989; Semikhatov *et al.*, 1991).

STRUCTURE OF THE FORMATION

The Osor-Khayata Formation within the area described crops out in a NE-trending band of 3 - 6 km in width and more than 30 km in length. The formation in this band forms a gently (2° - 7°) northwest-dipping monocline, separated by transverse faults into a series of blocks partially overlain by Permian deposits (Fig. 1). The faults are associated with sills and dikes of pre-Vendian basites and with wide, elongated, NW-trending zones that show a series of linear positive anomalies of magnetic field ΔTa (Fig. 1). The axes of these anomalies reflect tectonic disturbances that control a series of linearly elongated diabase dikes, most of which are not exposed at the surface.

Complexly-dislocated metamorphic rocks of the Eekit Formation of Early Proterozoic age and granitoids that penetrate them serve as the base of the Osor-Khayata Formation. The contact of gently deepening non-metamorphosed conglomerates and gravelly sandstones of the Osor-Khayata Formation with these rocks can be seen in the upper reaches of the Ortoku-Eekit River at the upper waterfall and in the midstream of the Burgaldzhi River, 8.5 km above its mouth.

The reference section of the formation at the Ortoku-Eekit River (Fig. 1, point 1) contains six members identified from prevailing varieties of silicoclastic rocks; their description is based on the study of natural outcrops, trenches, and partly of screes and eluvium debris. The following rocks occur at the upper reaches of this river at the upper waterfall, unconformably resting on the Early Proterozoic granitic pegmatites (in ascending order):

(1) Light grey (up to white) sandstones, which form coastal cliffs. The rock is predominantly quartz in composition (with varying admixtures of feldspars), mostly coarse- and medium-grained, with intercalations and lenses of gravelly varieties disappearing toward the top. The pebbles (up to 10 cm in size) are represented largely by quartz and microquartzite, from milky-white to black. They are, as a rule, angular, less frequently rounded, sometimes "floating" in a finer-grained groundmass. The texture is mosaic; the structure is massive and coarsely horizontally- and cross-laminated; the bedding surfaces show large ripple marks. The thickness is 4 m.

(2) Dark and light greenish- and brownish-grey, medium- and fine-grained sandstones with subordinated intercalations of siltstones. The rocks are horizontally- and cross-laminated, feldspar-quartz and quartz in composition, to various degrees glauconite-bearing (from 1 to 15 - 20%), micaceous and ferruginate in some beds, with porous and filmy (argillaceous, micaceous, glauconitic, and chloritic) cement. Quartz, feldspars, micaceous and titanium (leucoxene, anatase) minerals, and chlorite occur in the rocks as neogenic minerals. Chlorite replaces glauconitic globules and

Fig. 1. *Geological map of the Sololi-Sygynakhtakh Interfluve.*

1 – Lower Proterozoic metamorphic rocks and granites; 2 - 4 – Riphean deposits: 2 – silicoclastic rocks of the Osor-Khayata Formation, 3 – silicoclastic rocks of the Sygynakhtakh Formation, 4 – carbonate deposits of the Kyutingda Formation; 5, 6 – deposits of the Arymas Formation: 5 – silicoclastic-carbonate deposits, 6 – trachybasic lavas and their tuffs; 7 – Permian deposits; 8 – pre-Vendian sills, dykes and stocks of basic composition; 9 – faults; 10 – positive anomalies of ΔTa in the aeromagnetic map; 11 – sampling sites.

greenish-brown mica flakes, on which titanium and ferruginous minerals often occur. The glauconite-bearing varieties of cross-laminated feldspar-quartz fine-grained sandstones and siltstones in the lower part of the member are represented by sample 561b, described below. The thickness is 25 - 30 m.

(3) Light pinkish- and greenish-grey, dark-grey variously grained, horizontally- and cross-laminated, quartz-feldspar and quartz sandstones. Relatively coarse quartz material forms independent interbeds, lenses and portions of irregular shape or occurs as separate dark grains irregularly distributed ("floating") in a lighter fine-grained quartz-feldspar mass. The character of cementation is similar to that mentioned above. The rocks are enriched in glauconite, whose content varies from layer to layer; its relatively large grains (up to 0.4 mm) occur in the top of the member, where samples 560/2 and 560/1 from thin-laminated fine-grained feldspar-quartz sandstones were taken. The thickness is 4 m.

(4) Pinkish-, yellowish-, and greenish-grey coarse- and fine-grained (occasionally coarse-grained) quartz-feldspar and feldspar-quartz, coarse-laminated, massive sandstones with subordinated interbeds of finer-grained thin-laminated sandstones and siltstones. All rocks are enriched in rather large (up to 1.0 mm) glauconite grains. These rocks are represented in our collection by samples 560/3 and 560/5, taken several meters from one another along the section at the top of the member. Sample 560/5 was taken in block clitters at the same location on Mt. Ortoku-Eekit, where A.A. Krasil'shchikov had earlier taken a sample whose mineralogically-unstudied glauconite yielded a K-Ar age of 1435 Ma (Geokhronologiya Dokembriya ... , 1968). Glauconite globules, whose content varies in the samples studied from a few percent (sample 560/5) to 15 - 20% (560/3), occur both in coarse-grained feldspar-quartz sandstones with a mosaic texture and massive structure and in variously-grained sandstones. The latter are represented by coarse- to fine-grained varieties, in which large quartz grains float in a fine-grained quartz-feldspar matrix. The thickness is 8 m.

(5) Light-grey, beige, brownish-grey, predominantly fine-grained feldspar-quartz, quartz-feldspar sandstones and prevailing siltstones with an insignificant admixture of fine glauconite grain and varying concentrations of clayey, micaceous, and ferruginous minerals. The texture is similar to that of the above-mentioned fine-grained rocks; the structure is mainly laminated. The thickness is about 50 m.

(6) Greenish-grey, dark-grey to black (cherry near the top), silicified in interbeds and ferruginete fine-grained sandstones and siltstones. The rock is composed of quartz with varying admixtures of feldspars, small glauconite grains, micaceous, clayey and ferruginous minerals. The upper (predominantly silty) part of the member contains thin beds and lenses of brownish-grey, coarse-grained, sometimes stromatolitic dolomites. The thickness is 30 - 32 m.

The above rocks, whose total thickness is 125 - 130 m, are overlain by light- and pinkish-grey, medium-grained to cobble conglomerates of the Sygynakhtakh Formation, which contain the pebbles underlying Osor-Khayata rocks.

When tracing the contact of the Osor-Khayata and Sygynakhtakh formations northeast of the Ortoku-Eekit River, it is evident that the Sygynakhtakh conglomerates rest on different horizons of the Osor-Khayata deposits: on the coarse- to medium-grained sandstones of the top of the fourth member on the Burgaldzhi River, 6 km from the mouth, and on the fine-grained cream-colored sandstones and siltstones of the fifth member at the watershed of the Burgaldzhi and Sololi rivers. As a result, the thickness of the Osor-Khayata deposits remaining after the pre-Sygynakhtakh erosion varies from 130 to 50 - 60 m.

The structure and composition of the Osor-Khayata Formation, 15 km northeast of the Ortoku-Eekit River, about mid-stream of the Burgaldzhi River, remain similar to that described above, but fine-pebble conglomerates and gritstones appear here in the base of the formation. These rocks are light-grey and pinkish-yellowish, with quartz and less frequent feldspar pebbles 0.2 to 5 cm in size, variously-grained feldspar-quartz cement, and lenses of sandstones of a similar composition. Two meters up the section, the conglomerates and gritstones sharply decrease in quantity and remain only as lenses (up to 0.5 m in thickness) among light-grey, variously-grained feldspar-quartz sandstones with floating quartz pebbles. The appearance of conglomerates among the rocks of the first member is accompanied by its increasing thickness, up to 8 m. Higher horizons of the formation exposed along the Burgaldzhi River do not essentially differ in structure, composition, and thickness from those described for the Ortoku-Eekit River, except that among the sandstones of the fifth member, thin (0.1 - 0.2 m) lenses of quartz fine-pebble conglomerates and gritstones appear; the sixth member is partly cut off by the pre-Sygynakhtakh erosion; and thin interbeds of dark-grey, lilac, and cherry argillites are seen among its various-colored sandstones and siltstones.

The samples of glauconite-bearing rocks were taken in the section on the Burgaldzhi River from the sandstone-siltstone deposits at the base of the second member (samples 559/2, 051) and from the sandstones at the top of the fourth member (samples 2076, 2076A, and 558) (Fig. 1, points 2 and 3).

MINERALOGIC FEATURES OF FINELY DISPERSED, GLOBULAR AND LAMELLAR LAYER SILICATES

The finely dispersed layer silicates (fractions <0.6; <1 μm) studied in the glauconite-bearing samples that were collected at different stratigraphic levels (the second, third, fourth, and sixth members of the formation) in the two studied sections proved to be similar in mineral composition. According to X-ray and electron dif-

Table 1. Electron diffraction data for mica and chlorite (fractions <0.6, <1.0 μm)

Sample no.	Size, μm	Mica				Chlorite
		a, Å	*b*, Å	*c*, Å	β, deg	*b*, Å
560/5	<1	5.21	9.03	10.13	101.38	–
560/3	<1	5.21	9.02	10.31	101.49	9.29
560/2	<0.6	5.21	9.02	10.21	101.60	–
560/2	<1	5.21	9.02	10.19	101.46	9.28
558	<0.6	5.20	9.01	10.13	101.18	–
	<1	5.21	9.03	10.17	101.50	–
559/2	<1	5.21	9.03	–	–	9.29

Table 2. Electron diffraction data for the samples of Al-glauconite

Sample no.	Size, mm	Density, g/cm^3	*a*, Å	*b*, Å	*c*, Å	β, deg
Ortoku-Eekit River						
560/5	1 - 0.63	2.65 - 2.75	5.23	9.05	10.15	101.5
560/3	1 - 0.4	2.65 - 2.75	5.21	9.03	10.18	101.7
560/2	0.63 - 0.315	2.65 - 2.75	5.22	9.04	10.17	101.5
560/1	0.63 - 0.315	2.6 - 2.7	5.22	9.04	10.17	101.5
561b	0.315 - 0.1	–	5.22	9.04	10.18	101.3
Burgaldzhi River						
558	0.63 - 0.4	2.65 - 2.7	5.22	9.05	10.17	101.6
2076	0.63 - 0.4	2.65 - 2.7	5.22	9.04	10.19	101.16
2076A	1 - 0.4	2.65 - 2.75	5.23	9.06	10.18	101.07
2076A	0.63 - 0.4	2.65 - 2.75	5.23	9.05	10.18	101.5
051	0.4 - 0.16	2.7 - 2.75	5.23	9.05	10.16	101.4
559/2	0.315 - 0.16	–	5.22	9.05	10.22	101.4

fractometry data, they are represented by poorly crystallized dioctahedral mica of Al-composition, whose elementary cell has the parameter *b* equal to 9.01 - 9.03 Å (Table 1). Poorly crystallized trioctahedral FeMg chlorite (b = 9.28; 9.29 Å) and quartz, albite, and orthoclase are present as admixtures in the fraction <1 μm in some samples.

The grains of layer silicates were separated by the technique described by Semikhatov *et al.* (1987) and studied by a set of contemporary chemical and physical methods.

The glauconite grains in the studied samples are of globular (oval, round, kidney-shaped, and pisolitic), tabular (lamellar), and irregular shape. The globular shape more frequently occurs in grains confined to coarser-grained rocks (samples 560/3, 560/5, 2076, 2076A, and 558), whereas grains of irregular shape (samples 561b, 559/2, 560/1, 560/2, and 051) prevail in finer-grained varieties. The irregular shapes are caused by the interaction of the globules with the enclosing minerals, their deformation producing peculiar festoons and cement mass. The mineral in the festoons and cement mass, compared to undeformed globules, often exhibits poor pleochroism, longitudinal jointing, undulatory or parallel extinction, and increased interference colors. The grains of irregular festoon shape prevail in samples 560/1 and 560/2, where tabular varieties are also widely developed, especially in the latter sample (Fig. 2). These varieties exhibit most clearly high pleochroism, parallel extinction, and increased interference colors. Lamellar and festoon grain shapes are much more rare in samples 560/3, 558, 559/2, and 051, and they are completely absent in samples 560/5, 2076, and 2076A. It should be noted that the grains of lamellar and festoon shape more often have a glassy luster and an uneven surface, whereas the globular grains more often have a dull luster and a smoother surface. However, the surface of both kinds of grains can be cracked.

Based of the structural features of the grains studied, one can identify homogeneous and heterogeneous varieties with fine-aggregated and/or oriented textures among them. Uniformly-colored grains with no inclusions are rather rare in the samples; more often, one can see irregularly (with spots, veinlets, and zones) colored grains and/or grains with inclusions of various minerals. Tabular grains (Fig. 2) found in sample 560/2 (up to 80%) and in sample 560/1 (up to 40%) are often characterized by a combination of lamellar (oriented) and microaggregate textures. Microaggregate glauconite is often present between the cleavage faces of micaceous flakes that also belong to glauconite by composition, but have kept the shape of detrital mica, which is deformed, with uneven, sometimes fan-like contours.

The grain color is, on the whole, green, but its intensity and shades may vary somewhat, both in the same sample and from one sample to another. Thus, grass color is typical of samples 561b, 560/1, and 560/2, whereas bluish-green color occurs in others. Grains with a light (light bluish-green to whitish) central part and a dark bluish-green rim are typical of samples 2076 and 2076A and, to a lesser extent, of samples 558, 560/5, and 560/3 (Fig. 3a, 3b). The marginal rim of a grain can partially break off like a shell in the process of separation and open a light central zone. The intensity of lightness of globules in the same sample is different, however; one can find among them all transitional varieties from well-defined zonal globules to grains with a hardly noticeable light spot in the central zone. The microaggregate extinction of microcrystals is replaced by one that is partly oriented in the lighter parts, and the interference color becomes somewhat higher.

A study of the globules in a scanning electron microscope has revealed the domination of a lamellar nanostructure, typical for glauconite, which is observed both in color-uniform and clearly zonal varieties. The tabular grains typically have a combination of parallel

laminated and (between the laminae) lamellar structures.

The globule size varies from 1.0 to 0.1 mm on the average, coarse-grained varieties prevailing in samples 560/5 (1 - 0.63 mm) and 2076, 2076A, 558, and 560/3 (1 - 0.315 mm), and finer-grained varieties (0.315 - 0.2 mm), in samples 560/1, 560/2, and 051, the smallest grains (0.2 - 0.1 mm) being typical of samples 559/2 and 561b. Note that for detailed mineralogic-crystallochemical studies, the largest grains were chosen, wherever possible, in all samples, and these grains were divided into fractions by density when the amount was sufficient.

The grain density in the samples is within the range 2.6 - 2.75 g/cm^3. The prevailing density fractions (75 - 80%) for some samples (560/1, 560/3, 2076, 2076A, and 558) are in the range 2.65 - 2.7 g/cm^3, whereas for others (560/2, 560/5, and 051), they shift toward a heavier fraction (2.7 - 2.75 g/cm^3); the lightest grains, of density 2.6 - 2.65 g/cm^3, occur in samples 558 and 560/1 only in a subordinate amount (20 - 25%).

The analysis of diffractograms for oriented preparations of the grains (samples 560/1, 560/2, 560/3, 560/5, 2076, 2076A, 558, and 051) showed that they are represented by dioctahedral micaceous minerals, which are essentially devoid of swelling layers (≤10%). Samples 051 and 560/2 contain traces of chlorite, in addition to mica.

Diffractional patterns of nonoriented preparations of the grains are characterized by relatively low modulations in the region of reflexes 02L and 11L. The degree of mica deficiency determined by the ratio of reflexes 112 and $\bar{1}12$ to the diffractional maximum with $d \approx 2.58$ Å is 0.50, according to Sakharov *et al.* (1990).

An electron diffraction analysis of the samples revealed a high degree of crystallization of dioctahedral micaceous minerals. Judging from the parameter b, which is equal to 9.03 - 9.06 Å, these minerals belong to Al-glauconite (Table 2). This composition of micaceous minerals characterizes samples represented both by purely globular varieties (560/3, 560/5, 2076, 2076A, and 558) and by a mixture of irregular and tabular grains (560/1 and 560/2). The marginal darker and inner lighter parts of Al-glauconite globules in unevenly-colored varieties were separately studied in sample 2076. The parameter b of dark-green micaceous minerals proved to equal 9.05 Å; that of light-green minerals was 9.03 Å; their average parameter b, as seen from Table 2, equals 9.04 Å.

In addition, green, brownish-green, and brown tabular minerals from heavier fractions (≥2.75 g/cm^3), whose content in the samples is negligibly low (≤1%) compared to the prevailing lighter fractions mentioned above, were also studied by electron diffraction analysis (Table 3) in samples 560/1 and 560/2. Brown tablets (sample 560/1) consist of almost equal amounts of a well-crystallized dioctahedral Al-mica, illite with b equal to 9.02 Å, and of a low-crystallized trioctahedral chlorite (b = 9.23 Å). The amount of chlorite decreases

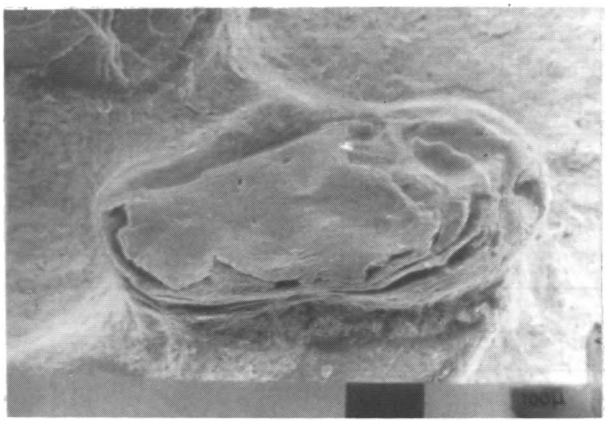

Fig. 2. *General view of Al-glauconite grains of tabular shape (Scanning electron microscope, a ×250 magnification).*

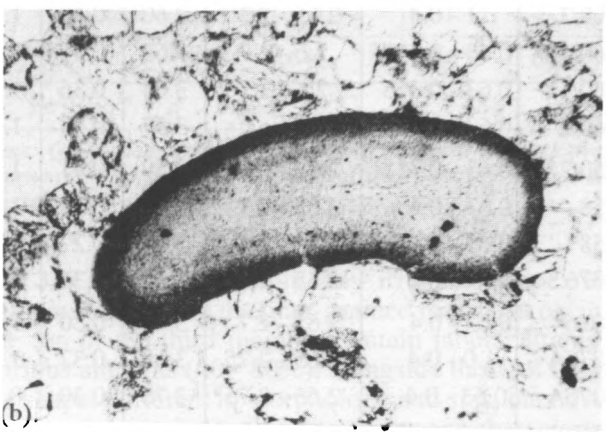

Fig. 3. *Al-glaugonite grains of globular shape: (a) general view (SEM, a ×100 magnification); (b) thinection (nic. 1, a ×100 magnification).*

in brownish-green tablets (sample 560/2); the mineral has varying composition (b = 9.24; 9.30 Å) and a poor degree of crystallization. The dioctahedral mica prevailing in the tablets is more ferruginous and belongs to Al-glauconite (b = 9.05 Å). The green tablets (samples

Table 3. Electron diffraction data for differently colored tabular grains

Sample no.	Size, mm	Density, g/cm^3	Color	Mica				Chlorite
				a, Å	*b*, Å	*c*, Å	β, deg	*b*, Å
560/2	0.63 - 0.315	2.75 - 2.8	Green	5.22	9.05	10.17	101.3	
560/2	0.63 - 0.315	2.75 - 2.8	Brownish-green	5.22	9.05	10.17	101.3	9.24; 9.30
560/1	0.63 - 0.315	≥2.75	Green	5.22	9.04	10.14	101.2	
560/1	0.63 - 0.315	≥2.75	Brown	5.21	9.02	10.18	101.2	9.23

Table 4. Chemical composition of the samples

Sample no.	Size, mm	Density, g/cm^3	SiO_2	Al_2O_3	Fe_2O_3	FeO	CaO	MgO	Na_2O	K_2O	H_2O^+	H_2O^-	Total
560/5	1 - 0.63	2.65 - 2.75	51.32	15.14	10.66	3.31	0.15	3.43	0.27	8.03	5.47	2.45	100.24
560/3	1 - 0.4	2.65 - 2.75	50.65	17.90	8.63	1.86	–	3.01	0.38	8.76	6.18	1.99	99.95
560/1	0.63 - 0.315	2.6 - 2.7	51.68	18.20	7.17	1.88	0.40	3.89	0.09	8.25	6.21	2.12	99.94
560/2	0.63 - 0.315	2.65 - 2.75	50.77	18.44	7.61	2.12	0.26	4.14	0.19	8.25	5.67	2.67	100.22
558	0.63 - 0.4	2.65 - 2.7	52.79	16.41	8.74	2.36	–	3.14	0.11	7.66	6.38	2.34	100.13
2076	0.63 - 0.4	2.65 - 2.7	52.06	15.88	10.06	2.40	0.04	3.33	0.11	7.66	6.34	1.99	99.86
2076A	0.63 - 0.4	2.65 - 2.75	51.62	14.50	11.19	3.07	0.35	2.65	0.34	7.30	6.20	2.04	99.28
2076A	1 - 0.4	2.65 - 2.75	52.21	16.30	10.35	1.56	0.08	3.27	0.11	7.68	6.45	1.82	99.99
051	0.4 - 0.16	2.7 - 2.75	51.80	17.98	8.76	3.14	0.30	2.42	0.30	7.26	6.24	1.53	99.74

Table 5. Crystallochemical features of the samples

Cations			Tetrahedral		Octahedral					Intralayer			*n*
Sample no.	Size, mm	Density, g/cm^3	Si	Al	Al	Fe^{3+}	Fe^{2+}	Mg	Σ	K	Ca	Na	
560/5	1 - 0.63	2.65 - 2.75	3.65	0.35	0.91	0.57	0.20	0.36	2.04	0.73	0.01	0.04	0.46
560/3	1 - 0.4	2.65 - 2.75	3.60	0.40	1.10	0.46	0.11	0.32	1.99	0.79	–	0.05	0.34
560/1	0.63 - 0.315	2.6 - 2.7	3.63	0.37	1.13	0.38	0.11	0.41	2.03	0.74	0.03	0.01	0.30
560/1*	0.63 - 0.315	2.6 - 2.7	3.64	0.36	1.15	0.30	0.19	0.41	2.05	0.74	0.03	0.01	0.30
560/2	0.63 - 0.315	2.65 - 2.75	3.57	0.43	1.10	0.40	0.12	0.43	2.05	0.74	0.02	0.03	0.32
560/2*	0.63 - 0.315	2.65 - 2.75	3.59	0.41	1.13	0.29	0.24	0.44	2.10	0.74	0.02	0.03	0.32
558	0.63 - 0.4	2.65 - 2.7	3.72	0.28	1.08	0.46	0.14	0.33	2.01	0.69	–	0.01	0.36
558*	0.63 - 0.4	2.65 - 2.7	3.73	0.27	1.10	0.41	0.20	0.33	2.04	0.69	–	0.01	0.27
2076	0.63 - 0.4	2.65 - 2.7	3.68	0.32	1.01	0.54	0.14	0.35	2.03	0.69	–	0.01	0.40
2076*	0.63 - 0.4	2.65 - 2.7	3.70	0.30	1.03	0.43	0.25	0.35	2.06	0.69	–	0.01	0.40
2076A	1.0 - 0.4	2.65 - 2.75	3.68	0.32	1.03	0.55	0.09	0.34	2.01	0.69	0.01	0.01	0.38
2076A	0.63 - 0.4	2.65 - 2.75	3.70	0.30	0.93	0.60	0.18	0.28	1.99	0.67	0.03	0.05	0.46
051	0.4 - 0.16	2.7 - 2.75	3.64	0.36	1.12	0.46	0.18	0.25	2.01	0.65	0.02	0.04	0.36

Note: Asterisk denotes samples whose formulas were recalculated with a more precise Fe^{2+}/Fe^{3+} ratio.

560/1 and 560/2) show no chlorite at all, and the dioctahedral mica (b = 9.04; 9.05 Å) also belongs to Al-glauconite (Table 3).

The averaged crystallochemical formulas were calculated only for the prevailing fractions in samples of green micaceous minerals based on complete silicate microanalyses (Tables 4 and 5), assuming a constant composition of the anion frame $(O_{10}(OH)_2)^{-22}$ for a half of the elementary cell. As seen from Table 5, aluminum prevails in the octahedral 2:1 lattices of mineral layers; its content varies from 0.91 to 1.15 formula units, with the ferrum degree (n = Fe^{3+} + Fe^{2+}/Fe^{3+} + Fe^{2+} + Al) varying from 0.30 to 0.46, which corresponds, according to Tsipurskii and Ivanovskaya (1988), to Al-glauc-

onite. The lowest Fe^{3+} content and the highest Mg content are found in samples 560/1 and 560/2, in which, as has been mentioned, tabular varieties occur in various quantities, along with grains of irregular shape.

To detect variations in the chemical composition of color-zonal globules, an electron microprobe analysis using a Camebax electron X-ray microanalyser was made on sample 2076 (analyst G.V. Karpova). The analysis showed that the lighter, central parts contain, compared to dark-green, marginal rims, a higher Al_2O_3 and lower Fe_2O_3 concentrations, but the concentrations of SiO_2, MgO, and K_2O remain similar in different zones. These ratios of Fe^{3+} and Al in the structure of different-colored micaceous minerals are also confirmed by electron diffraction results.

An IR-spectroscopic study of the samples was made in the region of stretching vibrations of OH groups, and the spectra were interpreted in accordance with the data of Dainyak *et al.* (1987). A stronger band with a maximum near 3575 - 3590 cm^{-1} (Al, Fe^{2+}; Al, Mg) and a less intensive band in the frequency range 3505 - 3525 cm^{-1} ($Fe^{2+}Fe^{3+}$; $Fe^{3+}Fe^{3+}$; $MgFe^{3+}$) are resolved in the minerals. The predominantly Al composition of the octahedral lattices of 2:1 mineral layers revealed by IR-spectroscopic methods is confirmed by diffraction and chemical data.

Mössbauer spectroscopy was made for four samples (560/1, 560/2, 558, and 2076), which permitted refinement of their crystallochemical formulas based on precision data for the Fe^{2+}/Fe^{3+} ratios (Tables 5 and 6). It is because of these studies that we have revealed, on the one hand, lower values of the Fe^{2+}/Fe^{3+} ratio in the samples (558, 2076) in which tabular Al-glauconite grains are essentially absent, and, on the other hand, an increase in this ratio with increasing concentrations of such grains from sample 560/1 (≈40%) to sample 560/2 (≈80%). It has thus been found that the presence of tabular Al-glauconite grains controls the Fe^{2+} concentration in the minerals.

Table 6. Mössbauer parameters of the samples

Sample no.		560/1	560/2	558	2076
Grain size, mm		0.63 - 0.315	0.63 - 0.315	0.63 - 0.4	0.63 - 0.4
Grain density, g/cm³		2.6 - 2.7	2.65 - 2.75	2.65 - 2.7	2.65 - 2.7
Fe^{3+}	*I*, %	21	27	33	31
	H, mm/s	0.31	0.33	0.37	0.32
	QS, mm/s	0.28	0.29	0.27	0.28
	CS, mm/s	0.64	0.66	0.65	0.65
Fe^{3+}	*I*, %	35	23	28	25
	H, mm/s	0.41	0.32	0.36	0.36
	QS, mm/s	0.51	0.55	0.51	0.51
	CS, mm/s	0.67	0.66	0.67	0.67
$Fe^{2.5+}$	*I*, %	9	9	13	15
	H, mm/s	0.71	0.44	0.65	0.77
	QS, mm/s	1.14	0.78	0.95	0.91
	CS, mm/s	0.80	0.88	0.88	0.93
Fe^{2+}	*I*, %	20	20	19	20
	H, mm/s	0.72	0.63	0.72	0.73
	QS, mm/s	2.02	2.13	2.03	1.86
	CS, mm/s	1.28	1.26	1.24	1.29
Fe^{2+}	*I*, %	15	21	7	9
	H, mm/s	0.35	0.34	0.27	0.33
	QS, mm/s	2.85	2.79	2.80	2.80
	CS, mm/s	1.39	1.42	1.43	1.42
	ξ^2	497	641	504	686
Fe^{2+}/Fe^{3+}		0.65	0.83	0.48	0.57

Note: QS – quadrupole splitting; CS – chemical shift relative to natrium nitroprusside; H – halfwidth of absorption line; I – intensity, ξ – reliability index.

CONCLUSION

On the southern slope and in the central part of the Olenek uplift, the lower part of the Lower Riphean section consists of relatively fine-grained silicoclastic deposits of the Osor-Khayata Formation, which are separated from the overlying sandy-conglomerate sequences of the Sygynakhtakh Formation by a clear erosional surface. Because of the different amplitudes of the pre-Sygynakhtakh erosion, the thickness of the Osor-Khayata deposits varies from 50 to 130 m. The deposits are mainly represented by glauconite-bearing sandstones and siltstones; rudaceous varieties (conglomerate and gravelstone) are subordinate, do not contain glauconite grains, and are mainly developed in the lower 4 - 8 m of the section.

The glauconite widely occurring in the formation is represented in relatively coarse-grained sandstones by large (1.0 - 0.315 mm) globules, and in siltstones and fine-grained sandstones, by smaller grains (≤0.315 mm) of predominantly irregular shape. Two samples taken from the finer-grained rocks that occur at the top of the third member contain tabular grains in various amounts (40 - 80%), alongside those of irregular shape. Different in morphology and size, but having similar density properties, the grains are characterized by similar textural-crystallochemical features. They are represented by the Al-variety of glauconite, which essentially does not contain swelling layers and has a rather high degree of order in the structure.

Globular and tabular Al-glauconites have formed at the stage of diagenesis, but in different ways. The globular glauconite seems to have formed by a synthesis from colloidal solutions, and the tabular glauconite, by a transformation of flakes of terrigenous Fe-mica (biotite). The latter process was accompanied by a split-

ting of the scales and a crystallization of a common microaggregate glauconite with a lamellar nanostructure, rather than lamellar-jointed one. The formation of tabular Al-glauconite grains on biotite by transformation, rather than by synthesis, method is proven by their morphologic and textural-crystallochemical features. The latter, in particular, exhibit increased Fe^{2+} and Mg content in such grains.

The transformation of trioctahedral biotite into dioctahedral Al-glauconite seems to have occurred in stages, judging from the composition of different-colored lamellar micaceous minerals in glauconite-containing samples. The content of trioctahedral chlorite rapidly decreases (from ≈50% down to traces) from brown, to brownish-green, and then to green varieties, whereas dioctahedral mica becomes more ferruginous (from illite to Al-glauconite). The timing of completion of the transformational processes in mica is difficult to estimate, and it is possible that these processes could take place alongside with katagenetic transformations. As is known (Kossovskaya and Drits, 1971, 1975), the transformation of trioctahedral Fe-mica (biotite) into dioctahedral mica and trioctahedral chlorite widely occurs precisely at this stage.

Katagenetic processes in silicoclastic deposits of the Osor-Khayata Formation manifested themselves unevenly. They were expressed in finer-grained interbeds as the deformation of glauconite grains of any morphologic type, and in coarser-grained interbeds, as the changes in the chemical composition of globules due to the replacement of Fe^{3+} by Al in the octahedral positions of the mineral, which resulted in its discoloration (clarification). The appearance of a light-green core in some globules is related to the occurrence of cracks that resulted from the breaking of their outer dark-green shell under loading. The assumption of katagenetic change in the Al-glauconite composition in different-colored zones of a globule does not contradict the isotope-geochronologic data obtained for these zones.

REFERENCES

Bibikova, E.V., Lobach-Zhuchenko, S.B., Semikhatov, M.A., *et al.*, Geochronologic Scale of the Precambrian for the East European Platform and Its Framing, *Izv. Akad. Nauk SSSR. Ser. Geol.*, 1989, no. 4, pp. 8 - 22.

Biterman, I.M. and Gorshkova, E.R., New Data on the Stratigraphy of Sinian Deposits in the Olenek and Kuoi-Daldyn Uplifts, *Soveshchanie po Stratigrafii Otlozhenii Pozdnego Dokembriya Sibiri i Dal'nego Vostoka* (Conference on the Stratigraphy of Late Precambrian Deposits of Siberia and Far East), Nauka, Novosibirsk, 1962, pp. 11 - 12.

Dainyak, L.G., Besson, G., Slonimskaya, M.V., *et al.*, Interpretation of IR-spectra of Dioctahedral Mica in the Region of Stretching Vibrations of OH Groups, *Mineral. Zh.*, 1987, vol. 9, no. 1, pp. 46 - 54.

Geokhronologiya Dokembriya Sibirskoi Platformy i Ee Skladchatogo Obramleniya (Precambrian Geochronology of the Siberian Platform and Its Folded Framing), Nauka, Leningrad, 1968.

Kats, A.G. and Florova, Z.B., New Data on the Stratigraphy of the Upper Proterozoic on the Southern Slope of the Olenek Uplift, *Pozdnii Dokembrii i Rannii Paleozoi Sibiri. Sibirskaya Platforma i Vneshnyaya Zona Altae-Sayanskoi Skladchatoi Oblasti* (Late Precambrian and Early Paleozoic of Siberia. Siberian Platform and the Outer Zone of the Altai-Sayany Fold Region), IGiG SO Akad. Nauk SSSR, Novosibirsk, 1986, pp. 65 - 84.

Komar, V.A., *Stromatolity Verkhnedokembriiskikh Otlozhenii Severa Sibirskoi Platformy i Ikh Stratigraficheskoe Znachenie* (Stromatoliths of Upper Precambrian Deposits of the Northern Siberian Platform and Their Stratigraphic Significance), Nauka, Moscow, 1966.

Kossovskaya, A.G. and Drits, V.A., Crystallochemistry of Dioctahedral Mica, Chlorites, and Corrensites as Indicators of Geologic Environments, *Kristallokhimiya Mineralov i Geologicheskie Problemy* (Crystallochemistry of Minerals and Geological Problems), Nauka, Moscow, 1975, pp. 60 - 69.

Kossovskaya, A.G. and Drits, V.A., *Voprosy Kristallokhimicheskoi i Geneticheskoi Klassifikatsii Mineralov Osadochnykh Porod* (Problems of Crystallochemical and Genetic Classification of Minerals of Sedimentary Rocks), Nauka, Moscow, 1971, pp. 71 - 95.

Sakharov, B.A., Besson, G., Drits, V.A., Kameneva, M.Yu., and Salyn', A.L., A Study of the Nature of Packing Defects in the Structure of Glauconites by the Method of Diffractional Profile Analysis, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1990, no. 12, pp. 97 - 109.

Semikhatov, M.A., Gorokhov, I.M., Ivanovskaya, T.A., *et al.*, The Rb-Sr and K-Ar Age of the USSR Riphean and Cambrian Globular Layer Silicates: Materials for Geochronometer Estimation, *Litol. Polezn. Iskop.*, 1987, no. 5, pp. 78 - 96.

Semikhatov, M.A., Shurkin, K.A., Aksenov, E.M., *et al.*, New Stratigraphic Scale for the USSR Precambrian, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1991, no. 3, pp. 3 - 13.

Shpunt, B.R., Shapovalova, I.G., and Shamshina, E.A., *Pozdnii Dokembrii Severa Sibirskoi Platformy* (Late Precambrian of the North Siberian Platform), Nauka, Novosibirsk, 1982.

Tsipurskii, S.I. and Ivanovskaya, T.A., Crystallochemistry of Globular Layer Silicates, *Litol. Polezn. Iskop.*, 1988, no. 1, pp. 41 - 49.

Yakshin, M.S., On the Lower Riphean Sygynakhtakh and Osor-Khayata Formations of the Olenek Uplift, *Pozdnii Dokembrii i Rannii Paleozoi Sibiri. Rifei i Vend*, IGiG SO Akad. Nauk SSSR, Novosibirsk, 1988, pp. 36 - 46.

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Seismostratigraphic Analysis of the Ninetyeast Ridge Sedimentary Cover at 10° - 18° S

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Abstract – Results of seismostratigraphic analysis of the continuous seismic profiling (CSP) data are submitted, which was obtained on the cruise of the R/V *Issledovatel'* to the Mascarene-Australian geotraverse. Two structural units were revealed at the Ninetyeast Ridge and Osborne Knoll. Shallow-water and subaerial facies were found, and wide-spread upthrows were registered in the lower Paleocene-lower Eocene sequence. In the upper younger (post-early Eocene) unit, pelagic sediments were recorded. The early Cenozoic history of the Osborne Knoll is closely related to that of the Ninetyeast Ridge.

The Ninetyeast Ridge is a major north-south lineament of the Indian Ocean, stretching more than 4500 km from 9° N to 32° S, approximately along 90° E (Geologo-Geophisicheskii Atlas ... , 1975). Until recently, the structure of its sedimentary cover, as compared with that of adjacent deep basins, was poorly studied. The central portion of the ridge evokes particular interest. It is bordered with two structures well expressed in bottom topography (the Osborne Knoll from the west, and the Ninetyeast Trench from the east).

The basement of the Ninetyeast Ridge is mainly composed of basaltic lavas that flowed out in subaerial conditions (Sites 214, 216, 253, 756, 757). The drilling results revealed a shallow-water stage in the Ridge evolution, with age ranging from the Late Cretaceous in the north to the Eocene in the south (von der Borch, Sclater, *et al.*, 1974; Leg 121 Traces ... , 1988).

Since 1986, "Yuzhmorgeologiya" has conducted the geological and geophysical investigations at the Trans-Indian (Mascarene-Australian) geotraverse. Geophysical investigations on the geotraverse include a reference network of WAM (wide-angle method) profiles, continuous seismic profiling (CSP) of the area with frequency of 70 - 130 Hz, magnitometry, and echosounding. This work presents the results of seismostratigraphic analysis of (CSP) data obtained in the course of the R/V *Issledovatel'* cruise to the Ninetyeast Ridge and adjoining parts of basins (Pilipenko, 1990). A part of the geotraverse described encompasses the Ninetyeast Ridge at 10° - 18° S (Fig. 1).

CSP was conducted using a four-channel piezoarray with a one-channel, 25-m-long base, and a distance of 50 m between them. Elastic waves were excited by an Impulse-1 pneumatic transducer. The distance between the points of elastic wave excitation was 100 m, which ensured a time record density (distance between neighbouring traces) of every 25 m when processing the four-channel seismograms. CSP time records were treated in the Computer Center (CC) on land, using modern processing graphs, including reverse filtration and migration; these appeared to be a good basis for seismostratigraphic interpretation and distinguishing the geodynamic types of tectonic dislocations.

The Ninetyeast Ridge has a pronounced block morphology (Udintsev, 1987). Along the strike, it is differentiated into a series of blocks 50 - 80 km long, separated by faults with an amplitude up to 0.7 - 1.0 km, which stretch mostly northeast-eastward and control large valleys in ocean floor and basement topography. The valleys have steep southern and gentle northern sides. Separate blocks are arranged en-echelon. Within the geotraverse, the ridge width varies along the strike. Its northern portion averages 100 km, and the southern one, 200 km. The minimum depth above the ridge crest is about 1600 m in its central part.

The Osborne Knoll adjoins the Ninetyeast Ridge from the west. According to the seismostratigraphic data, it is genetically related to the Ninetyeast Ridge formation and is separated from the ridge by a valley opening into the neighbouring Central Basin. Sedimentary cover in the described area is explored at DSDP Site 214 (von der Borch, Sclater, *et al.*, 1974) and ODP Site 757 (ODP Science Operator Report, 1988). According to paleontologists, the age of the basal sediments is determined as the late Paleocene.

The drilling at the sites revealed the sedimentary cover to consist of two different units. The lower one is composed of volcano-sedimentary and shallow-water carbonate-terrigenous (Paleocene-lower Eocene) deposits. The upper unit comprises carbonate ooze of post-early Eocene age. The average velocity of elastic waves in the sedimentary cover at Site 214 is 1.9 km/s (von der Borch, Sclater, *et al.*, 1974).

The seismostratigraphic analysis of CSP data shows that within the Ninetyeast Ridge and Osborne Knoll, the sedimentary cover consists of two seismic com-

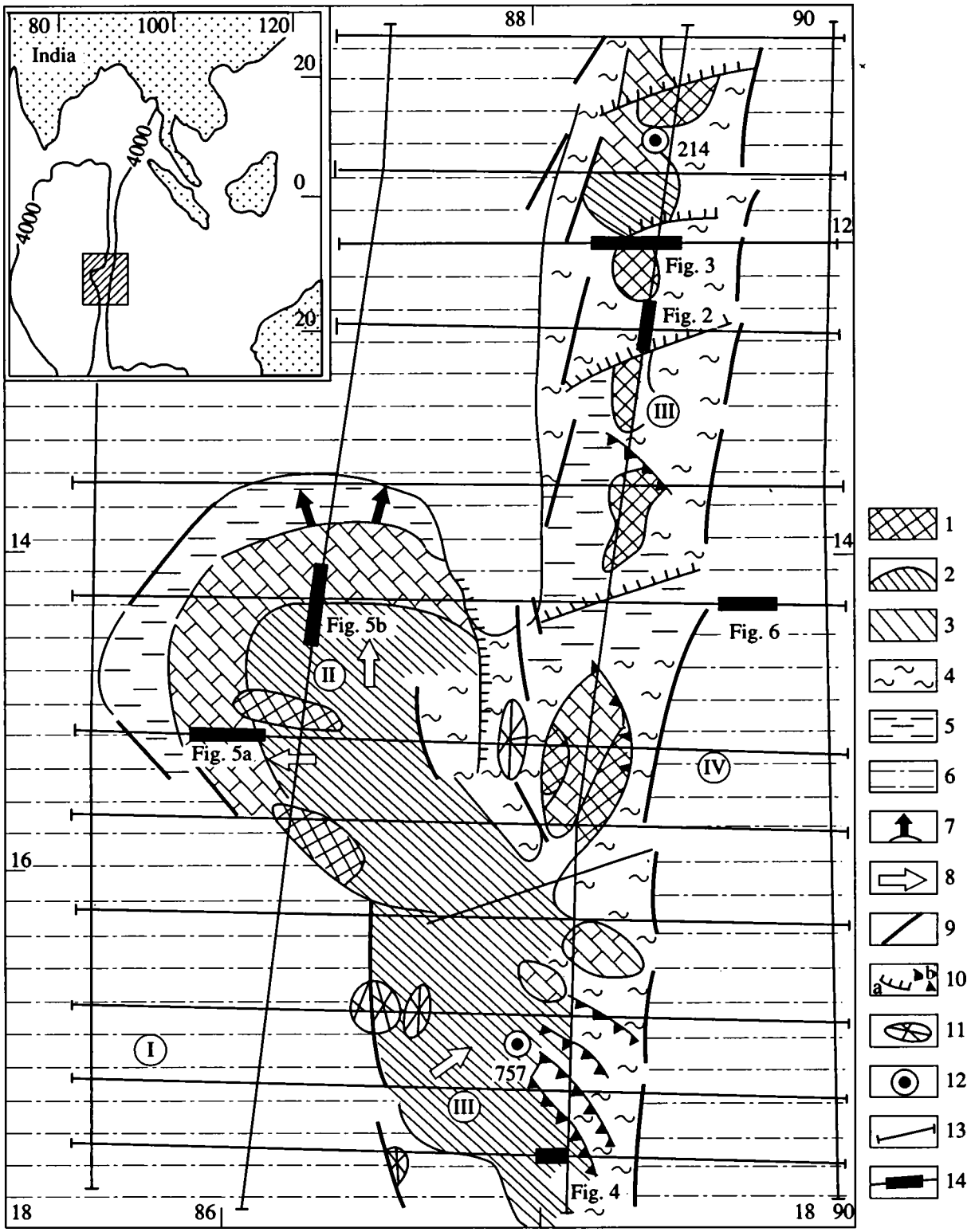

Fig. 1. *Scheme of seismofacies of Paleocene-lower Eocene complexes on the Ninetyeast Ridge and Osborne Knoll.*

1 – a zone devoid of Paleocene-lower Eocene sediments. Seismofacies: 2 – nonmarine; 3 – nonmarine and shallow-water; 4 – of the slope with chaotic filling; 5 – of slope with flat-parallel bedding; 6 – deep-water; 7 – base discordance in fondoform subzone of paleoslope; 8 – direction of inclination of Paleocene-lower Eocene sedimentation surfaces; 9 – main faults bordering the Ninetyeast Ridge and Osborne Knoll; 10 – rupture dislocations: faults (a), upthrows (b); 11 – volcanoes; 12 – drill sites; 13 – CSP profiles; 14 – fragments of profiles shown in Figs. 2 - 6. I – Central Basin, II – Osborne Knoll, III – Ninetyeast Ridge, IV – East Indian Trench. A small-scale insert shows the position of the area studied.

plexes (Figs. 2 - 5): the upper (SC-I) comprises pelagic sediments of post-early Eocene age, and the lower (SC-II) is represented by a complex combination and lateral transition of subaerial, shallow-water, and deep-water facies of Paleocene-early Eocene age.

Subdivision of the section into the upper (SC-I) and lower (SC-II) seismic complexes was made using seismostratigraphic parameters (Kunin, 1989; Seismic Stratigraphy, 1977; Shlezinger, 1989). When distinguishing subcomplexes and seismofacies, well-known seismostratigraphic criteria, such as reflection configuration, amplitude, extent of continuous tracing, apparent frequency, degree of transparency, and others, were used.

Seismic complexes SC-I and SC-II are separated by a regional unconformity inside the lower Eocene interval of the section. The mentioned unconformity surface is traced throughout the area and marks a stage of major tectonic reorganization in the eastern part of the Indian Ocean.

Tectonic fracturing of the ridge is clearly seen on the CSP profiles. Within the ridge, the basement rocks are broken into small blocks of 3 - 5 km and less. They are separated by tectonic dislocations (upthrows), displacing, as a rule, the whole sedimentary sequence of the lower complex (Fig. 2). Surfaces of the disruptions are inclined at an angle of 55° - 90°, with a tendency to flatten at the second layer level.

On profiles across the southern portion of the ridge, buried upthrow-thrusts with a 500 m amplitude are revealed, confined mostly to the lower sedimentary complex, and thus, not expressed in the present bottom topography. In some areas, the lower complex is deformed into a series of asymmetrical folds with an amplitude decreasing upward. In the northern portion of the ridge, the disjunctive dislocations diverging upwards are registered in the basement and sedimentary cover (Fig. 3). Similar dislocations are usually characteristic of the shear zone (Silvester, 1988). Thus, it can be inferred that major compressional deformations occurred in the Ninetyeast Ridge at an early stage of tectonic evolution (in the late Paleocene-early Eocene), and formation of the main unconformity between the two major sedimentary complexes was caused by tectonic activity.

Subaerial and shallow-water facies on the Ninetyeast Ridge are confined mainly to separate isolated uplifts (Fig. 1), which were elevated above sea level in the late Paleocene-early Eocene. In CSP profiles, seismofacies of nonmarine sediments are represented as an intermittent record with a sharp change in reflection amplitude. More intensive reflectors may correspond to lignite intercalations observed in the Site 214 section (von der Borch, Sclater, *et al.*, 1974). Such a form of the seismic record changes (sometimes with alternation) into a hummocky and chaotic or a small hummocky form of recording, the size of hummocks approaching the resolution limit of the method (50 - 70 Hz), which suggests a thin-layered character of the section. In extrapolating Site 757 data, one can suggest that this seismic facies corresponds to volcanic ashes and tuff (in Site 214 with lignite intercalations). Its thickness is 400 m.

Seismofacies in the form of continuously traced subparallel reflectors with high reflection amplitudes (seismofacies of strong reflection) are thought to be related to the shallow-water depositional environments. In some profiles, diverging reflectors are recorded, which can

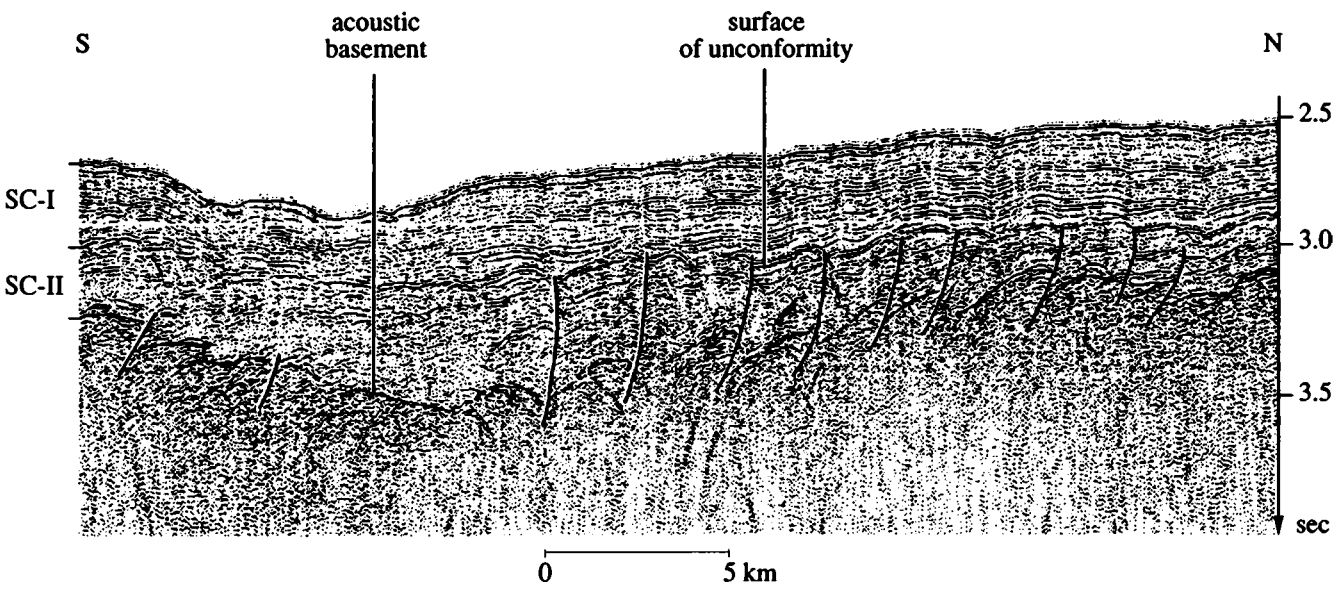

Fig. 2. *Fragment of seismic time section in CSP profile characterizing the structure of sedimentary cover and acoustic basement in the northern area, along the Ninetyeast Ridge strike (see Fig. 1 for the sequence location).*

The figure shows upthrows displacing the surface of acoustic basement and the lower sedimentary complex. SC – seismic complexes. Vertical scale – two-way travel time in seconds.

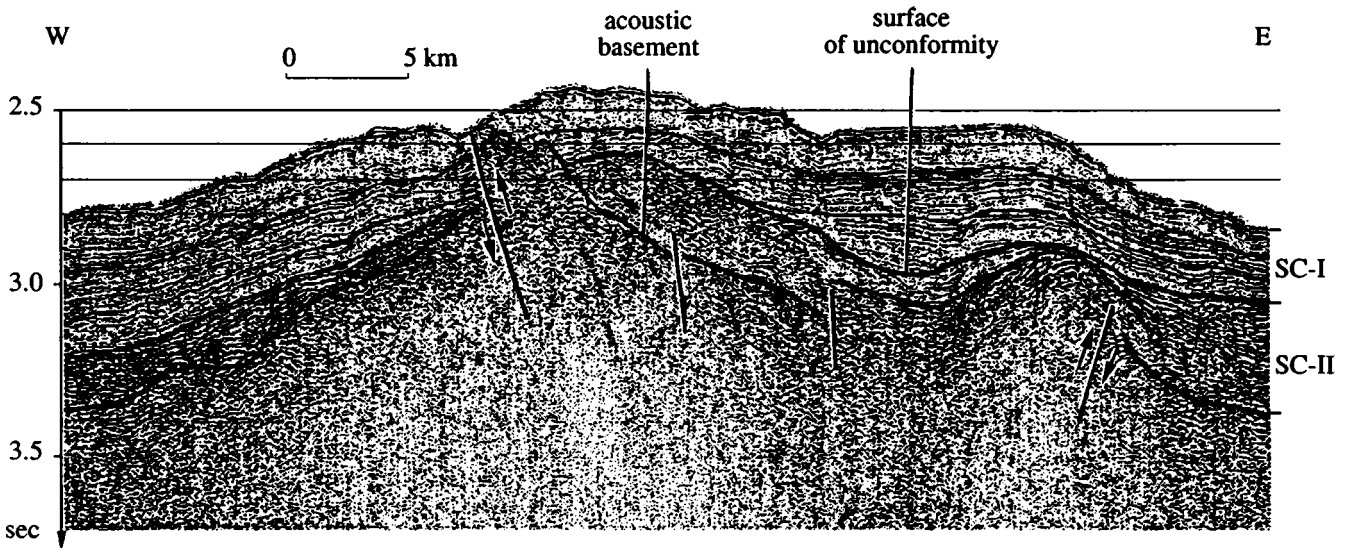

Fig. 3. *Fragment of seismic time section in CSP profile across the strike of the Ninetyeast Ridge (at 12° S).*

See Fig. 1 for sequence location. Vertical arrows show direction of block displacement along the supposed rupture dislocations. Vertical scale – two-way travel time in seconds. SC – seismic complexes.

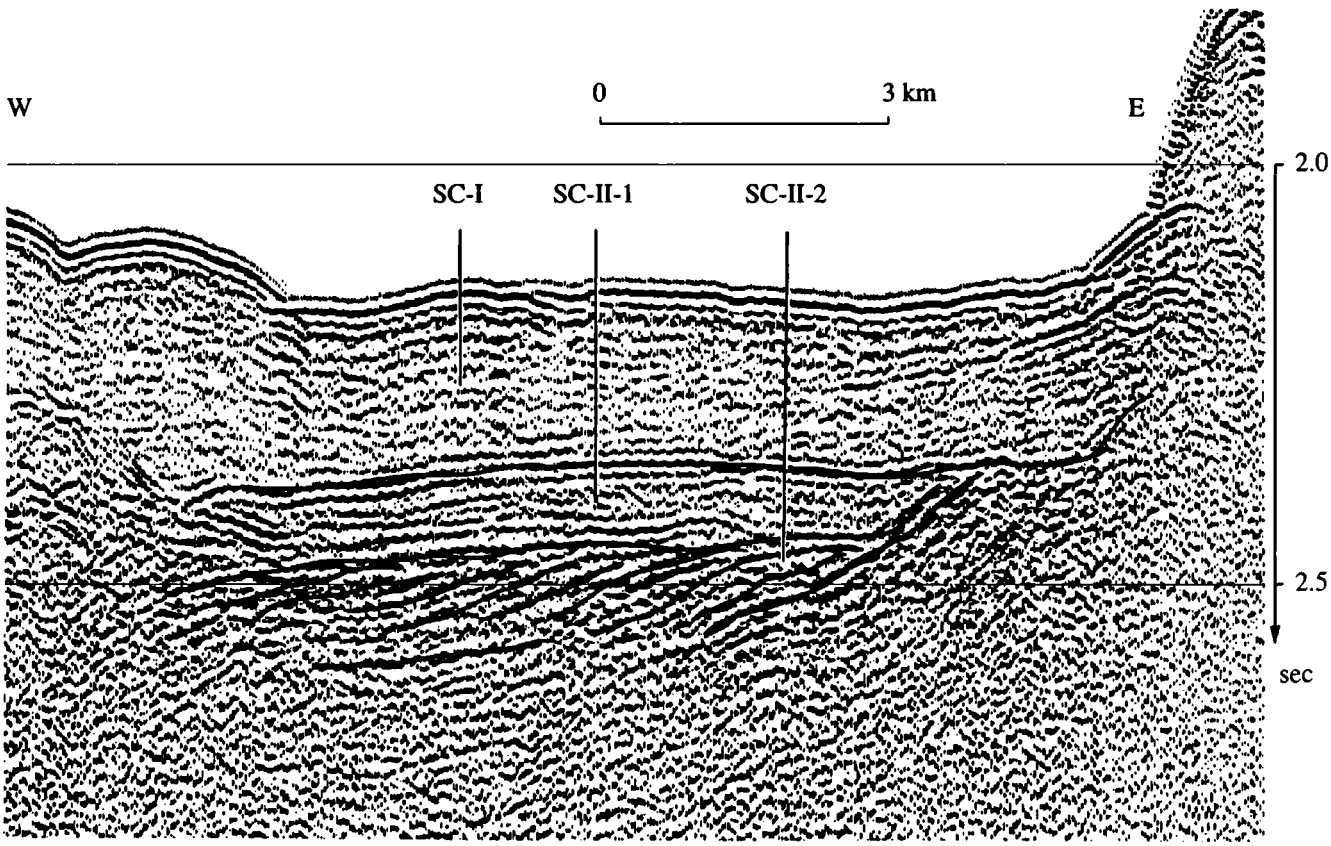

Fig. 4. *Fragment of seismic time section in CSP profile illustrating seismostratigraphic subdivision of sedimentary cover on the Ninetyeast Ridge top in the southern part of the area.*

The sequence location is shown in Fig. 1. SC – seismic complexes. Vertical scale – two-way travel time in seconds.

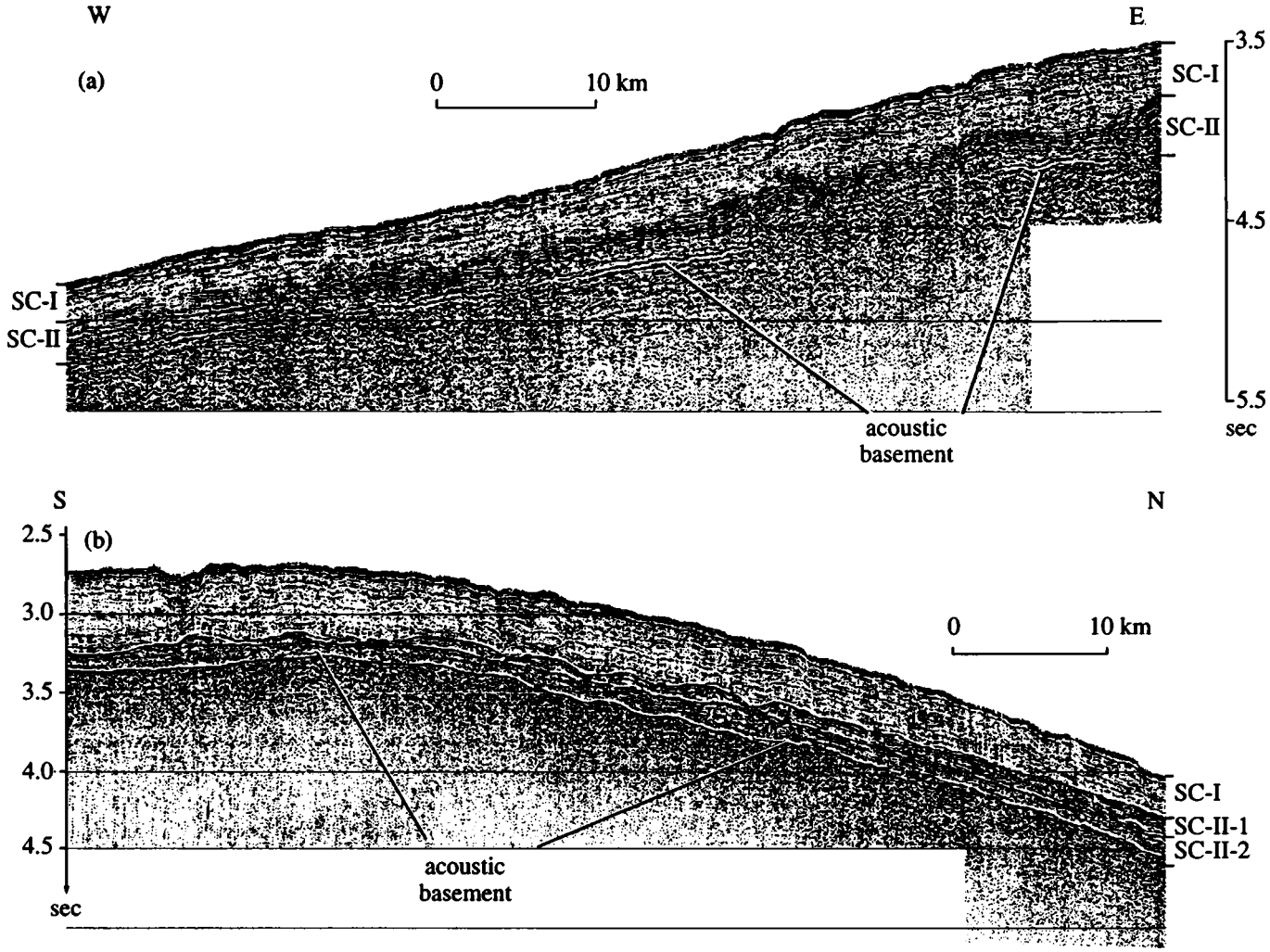

Fig. 5. *Fragments of seismic time sections illustrating seismostratigraphic subdivision of sedimentary cover onto seismic complexes within the Osborne Knoll in latitudinal (a) and submeridional (b) profiles of CSP.*

Location of sequences is shown in Fig. 1. SC – seismic complexes. Vertical scale – two-way travel time in seconds.

be interpreted as a gradual seaward inclination of the shelf sedimentation surface. In a lithological respect, the seismofacies of strong reflections are identified with shallow-water (the shelf) and deeper-water (the upper slope) carbonate rocks. In the shallow-to-deep-water transition area of the eastern slope of the Ninetyeast Ridge, a base discordance between SC-II reflectors and the basement surface (in the fondoform subzone of the paleoslope) is observed. East of the ridge top, grabenlike structures and flat terraces are recorded, which appeared to be sediment traps that were gradually filled with sedimentary, mainly volcaniclastic, material. Here, seismofacies change from weak to chaotic reflections, which is consistent with the depositional environment in paleoshelf-break zone. Accumulation of volcano-sedimentary facies of SC-II resulted in a flattening of the basement relief. The maximum thickness of the SC-II sediments in separate grabens and slope valleys amounts to 800 m.

In some areas of the southern Ninetyeast Ridge in SC-II, the CSP profiles revealed two subcomplexes, SC-II-1 and SC-II-2, separated by unconformity. In the lower subcomplex, SC-II-2 (above the basement), seismic facies of lateral accretion (oblique bedding) were registered (Fig. 4). The latter are interpreted as delta facies of late Paleocene (?) age.

A similar subdivision of the lower seismic complex into subcomplexes SC-II-1 and SC-II-2 is observed on the Osborne Knoll (Fig. 5). The seismofacies of subaerial and shallow-water origin are singled out here in SC-II. It was established that in the southwestern part of the Osborne Knoll, the boundary between the sedimentary cover and the acoustic basement (the base of the lower complex) is marked by an erosion surface. It is distinguished by a low-frequency multiphase record and is well expressed dynamically (Fig. 5a). The smooth pattern of the boundary is most likely indicative of onland erosion. Seismofacies sin-

gled out in the lower complex on the Osborne Knoll and in adjacent areas of the Central Basin are grouped quite regularly to form a peculiar lateral row (Fig. 1). Seismostratigraphic data indicates that the SC-II-1 and SC-II-2 sequences mentioned above were mainly formed in shallow-water (shelf carbonates) and continental conditions. From north to south (paleolandward), the thickness of SC-II-1 decreases gradually from about 100 - 150 m to zero values (Fig. 5b).

The accumulation of shallow-water carbonate sediments on the Osborne Knoll was evidently followed by a short-term sea-level fall. It is fixed by the presence of superimposed clynoforms at the foot of the slope and by a ramified net of a drainage (erosional) system in the SC-II sequence top (Fig. 5b). Local erosion is well observed in sublatitudinal profiles, i.e., across the strike of the SC-II sedimentation surface tilt.

The lower part of the paleoslope coincides with the current northern foot of the Osborne Knoll, where a base discordance is revealed between SC-II reflectors and the surface of the ocean basement.

The composition of sediments encountered at Sites 214 and 757 testifies that the upper seismic complex (SC-I) in the Ninetyeast Ridge and Osborne Knoll is composed of foraminifer-nannofossil and nannofossil ooze of middle Eocene-Pleistocene age. Its thickness varies from 100 to 450 m. The depositional environment corresponding to this complex was pelagic. Sedimentation was controlled by bottom currents, which is confirmed by a ramified system of buried minicanyons and zones of local accumulation. According to seismostratigraphic data, the Ninetyeast Ridge gradually subsided during the past 40 - 45 Ma. In some parts of the ridge, the lower strata of the SC-I section are absent in the stratigraphic sequence (Fig. 3). A base discordance is recorded here between SC-I reflectors and the deformed surface of the SC-II lower sedimentary complex, or the acoustic basement if the complex wedges out. Figure 3 shows that the upper seismofacies unit (SC-I), most likely of late Miocene age but no older, is traced continuously through the entire ridge, resting evenly on its top and slopes. The unit is characterized by a clearly transparent acoustic record.

Intensive tectonic dislocations are typical for the eastern foot of the Ninetyeast Ridge, including a conjugate side of the trench of the same name. A peculiar feature of the Ninetyeast Trench sedimentary cover is the presence of enveloping seismic complexes in the lower part and the lateral seismic complex filling up in the upper part. Formation of the first complex is related to pelagic sedimentation, while that of the second, to turbidity flow activity. It may be assumed that turbidites were formed as a result of carbonate material derived from uplifted adjacent parts of the Ninetyeast Ridge. In the trench, the postsedimentation inclinations of originally horizontal turbidite layers are revealed (Fig. 6). This suggests reactivating tectonic movements in the late Miocene-Quaternary at the eastern boundary of the Ninetyeast Ridge. High seismicity in the eastern side of the Ridge is evidence of the current tectonic movements (Petroy and Wiens, 1989).

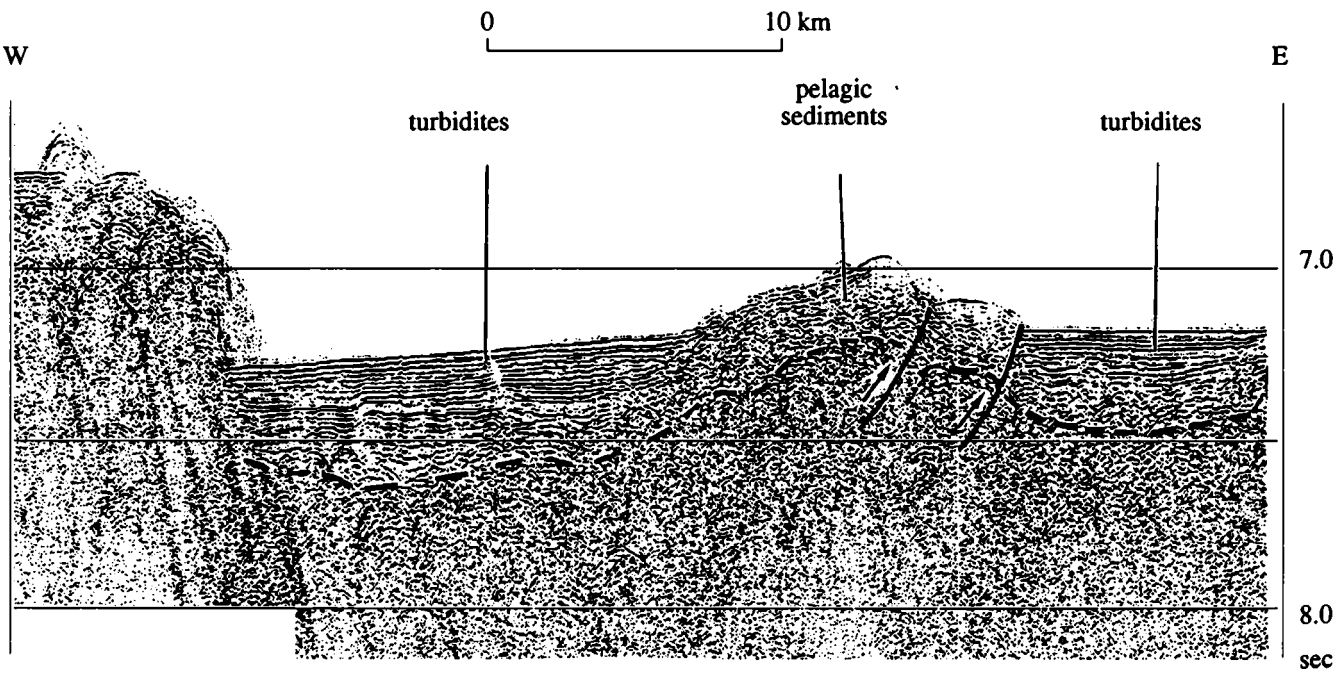

Fig. 6. *Fragment of seismic time section in CSP profile, illustrating postsedimentary inclination of sea bottom surface in the Ninetyeast Trench.*

See Fig. 1 for sequence location. Arrows show direction of upthrow block displacement along the rapture dislocations; dotted line shows the position of acoustic basement surface.

CONCLUSION

(1) The sedimentary cover of the Ninetyeast Ridge is composed of two structural units: the upper, composed of pelagic sediments of post-early Eocene age, and the lower, represented by a complex paragenesis of subaerial shallow-water and deep-water facies. One of the characteristic features of the late Paleocene-early Eocene sedimentation is the presence of a depositional system that evolved simultaneously on the Ninetyeast Ridge and Osborne Knoll in a continental and shallow-water environment and comprised carbonate, terrigenous, and volcanogenic sediments.

(2) A wide distribution of upthrow-thrust structures in the ocean basement and in the lower part of the sedimentary cover is new geological information obtained as a result of a CSP survey with a large (about 25 m) density of data points. Thus, intensive compression occurred at an early stage of tectonic evolution of the Ninetyeast Ridge. Tectonic compression in the early Eocene resulted in the formation of a regional unconformity at the boundary between two major sedimentary complexes within the Ninetyeast Ridge and the Osborne Knoll.

(3) The stratigraphic sequence of seismic complexes and seismofacies indicates a gradual subsidence of the ridge and Osborne Knoll in the Cenozoic. Processes of long-term compression played a significant role in the formation of the Ninetyeast Ridge as a regional morphostructure underwent several phases of tectonic activity (Korsakov and Pilipenko, 1991).

REFERENCES

Geologo-Geofizicheskii Atlas Indiiskogo Okeana (Geological and Geophysical Atlas of the Indian Ocean), Akad. Nauk SSSR, GUGK, Moscow, 1975.

Korsakov, O.D. and Pilipenko, A.I., Tectonic Deformations of the Ninetyeast Ridge, *Dokl. Akad. Nauk SSSR,* 1991, vol. 320, no. 2, p. 407 - 411.

Kunin, N.Ya., Sedimentation Models and Seismostratigraphy Insights, *Byull. MOIP. Otd. Geol.*, 1989, vol. 64, no. 1, pp. 24 - 32.

Leg 121 Traces Rifting and Hot Spots, *Geotimes*, 1988, vol. 33, no. 11, pp. 9 - 11.

ODP Science Operator Report. Leg 121: Broken Ridge and Ninetyeast Ridge Program, *JOIDES J.*, 1988, vol. 14, no. 3, pp. 5 - 8.

Petroy, P.E. and Wiens, D.A., Historical Seismicity and Implications for Diffuse Plate Convergence in the Northeast Indian Ocean, *J. Geophys. Res.*, 1989, vol. 94, no. 89, pp. 12301 - 12319.

Pilipenko, A.I., Structure and Early Cenozoic Evolution of the Ninetyeast Ridge (at 10° - 18° S), Abstracts of Papers, *9 Vsesoyuzn. Shkola Morskoi Geologii* (9th All-Union Workshop on Marine Geology), IO Akad. Nauk SSSR, Moscow, 1990, vol. 2, pp. 75 - 76.

Seismic Stratigraphy – Application to Hydrocarbon Exploration, Mem. 26 AAGP, Tulsa, Oklahoma, USA, 1977.

Shlezinger, A.E., Structure and Evolution of Sedimentary Basins by Seismostratigraphy Data, *Byull. MOIP. Otd. Geol.*, 1989, vol. 64, no. 1, pp. 33 - 38.

Silvester, A.C., Strike-Slip Faults, *Bull. Geol. Soc. Am.*, 1988, vol. 100, pp. 1666 - 1703.

Udintsev, G.B., *Rel'ef i Stroenie Dna Okeana* (Relief and Structure of Ocean Floor), Nedra, Moscow, 1987.

Von der Borch, C.C., Sclater, J.G., *et al.*, *Initial Reports of the Deep Sea Drilling Project*, vol. 22, U.S. Government Printing Office, Washington, 1974.

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BRIEF

COMMUNICATIONS

On the Geochemical Specialization of the Cretaceous-Paleogene Boundary "Iridium Horizon"

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In many regions of the world, where a continuous sequence of Cretaceous-Paleogene strata is preserved, a layer of clayey rock, enriched in iridium and a number of other elements, was recorded at the boundary of the Maestrichtian and Danian stages. The thickness of this layer does not exceed several centimeters. The transition from the Mesozoic to Cenozoic is considered to be accompanied by the rapid or, on a geological scale, almost instantaneous extinction of many species of flora and fauna. It is suggested that the formation of the *"iridium horizon"* and biotic changes are interrelated.

There are several hypotheses of the iridium anomaly origin at the Cretaceous-Paleogene boundary, two being the most well-grounded. The first explains the anomaly as a result of the Earth's collision with a large cosmic body and is based on: (1) high contents of iridium and other siderophile elements in the boundary deposits of the Cretaceous and Paleogene, (2) cosmic ratios of these elements, (3) a presence of impact-metamorphosed grains of quartz and feldspar, (4) a narrow stratigraphic localization of the geochemical anomaly, and (5) its global distribution (Alwarez *et al.*, 1980; Nazarov *et al.*, 1988). The second hypothesis considers the iridium horizon as a result of volcanic activity (Neruchev, 1988; Courtillot, 1990), which was exceptionally intense in the Late Cretaceous (Ronov *et al.*, 1989). Furthermore, according to recent data, the time of an outpouring of the Deccan traps (Courtillot, 1990), belonging to the largest lava fields of the world, corresponds exactly to the time of the iridium horizon formation at the Cretaceous-Paleogene boundary.

Lately, other notions have appeared explaining the iridium anomaly origin from a purely sedimentological position. Thus, D.P. Naidin (Naidin, 1986; Naidin *et al.*, 1990) considered high isidium contents to be a result of relative enrichment in the meteoritic component of an insoluble part of carbonate sediments, which were dissolved by increasing the carbon dioxide partial pressure in marine water. This increase resulted from the mass extinction of phytoplankton (the main consumer of CO_2) because of its poisoning by bacterial toxins during the gigantic "red flood."

V.A. Zakharov and A.S. Lapukhov (1991) consider the effect of sedimentation undercompensation (e.g., an abrupt decrease of the sedimentation rate against the background of a uniform supply of cosmic dust) to be one of the possible reasons for the anomalously high contents of iridium and some other elements in the deposits.

We have attempted to specify the origin of the iridium horizon with geochemical methods. We compared the explored horizon with country rocks. The thickness of the underlying and overlying strata, yielding geochemical information, varied from 10 to 72 cm. Substantial diversity in the sets of chemical elements in sections from various regions, which sometimes hampers their comparison and the availability of only qualitative information for other sections, determined the sampling volume and the set of the elements for geochemical comparison.

We studied two data massifs with 13 analyses for each: the first, for the iridium horizon, and the second, for the country rocks. We used geochemical information from the New Zealand (Brooks *et al.*, 1986), Denmark (Alwarez *et al.*, 1980), and Turkmenistan (Alekseev *et al.*, 1988) sections. A distribution of six elements (Co, Cr, Cu, Ni, Zn, and Ir) was estimated in every case. We calculated the average contents, values of dispersions, and coefficients of pair correlation for these elements. The samples were compared by estimating the dispersion value distinctions and average contents of the elements, using Fisher's variance ratio (F) and Student t test (t). The results obtained (Table 1) show a distinct difference between the Cretaceous-Paleogene boundary clay stratum and the country rocks, according to the dispersion values and average contents (except Cu). The values obtained for F and t are noticeably higher than the standard ones at a significance level of 0.05.

An examination of the results of the correlation analysis (Table 2) and the schematics of the significant correlations between the trace elements from the iridium horizon and from the country rocks (Fig. 1) shows: (1) significant positive correlations among Ir, Cr, and Ni in the boundary clays (for Turkmenistan, the set of elements positively correlated with Ir is more extensive

Table 1. Mean contents and dispersions of some trace elements from the clays of the iridium horizon and from country rocks

Elements	Iridium-horizon clays		Country rocks of the iridium horizon		Value of the Fisher's variance ratio (F)	Value of Student t test (t)
	average contents	dispersion	average contents	dispersion		
Co	81.8	5096	21.8	481	10.6	2.7
Cr	142.2	6627	38.5	661	10.0	4.2
Cu	138.2	30,668	75.5	2583	11.9	1.2
Ni	266.8	114,365	33.9	419	272.8	2.4
Zn	476.2	275,010	183.1	30,405	9.0	1.8
Ir	25.0	900	0.05	1.27×10^{-2}	70,866.0	2.9

Note: Element contents of Co, Cr, Cu, Ni, and Zn are given in g/t; of Ir, in kg/t.

(Nazarov *et al.*, 1988)); (2) an absence of significant positive correlations of iridium with any elements and, furthermore, a negative correlation between Ir and Cr (with a probability of 0.1) in the country rocks. Apparently, there was a single source of iridium and the elements positively related to it (Cr and Ni), whereas in the country rocks, iridium is an alien element in enclosing rocks and is not associated with any single process.

This conclusion is confirmed by the results of factor analysis, a wide-spread multivariate method, which is used in different branches of geology (Krumbein and Graybill, 1965; Davis, 1975). The data obtained (Tables 3, 4) represents iridium as an indicator of some inessential factor in the formation of country rocks, which had no strong influence on other elements. Otherwise, formation of a geochemically individual iridium horizon was conditioned by some common and rather intense factor, which influenced the above-discussed elements (including iridium) unidirectionally.

Table 2. Results of correlative analysis between trace elements from the clays of the iridium horizon and from country rocks

Elements	Cr	Cu	Ni	Zn	Ir
Clays from the iridium horizon					
Co	0.17	0.36	0.31	0.40	0.16
Cr		0.01	0.86	0.17	0.57
Cu			0.03	0.76	−0.14
Ni				0.26	0.53
Zn					0.07
Country rocks					
Co	0.56	0.31	0.06	0.01	−0.19
Cr		−0.03	−0.07	−0.15	−0.44
Cu			0.45	0.79	0.07
Ni				0.40	−0.10
Zn					0.04

A comparative analysis of the geochemical features of the iridium horizon and country rocks testifies to their distinct geochemical specialization. This is enough to reject the versions of the sedimentological hypothesis, which cannot explain the geochemical differences of the compared objects.

The two other hypotheses, i.e., the cosmic and volcanic, are assumed (Courtillot, 1990) to agree with well-known geological data. However, only ascertaining the direct transition of the iridium horizon into impactites or other volcanic formations can unambiguously prove either of these hypotheses. As there is no evidence of such spatial and genetic connections, the pronounced geochemical individuality of the iridium

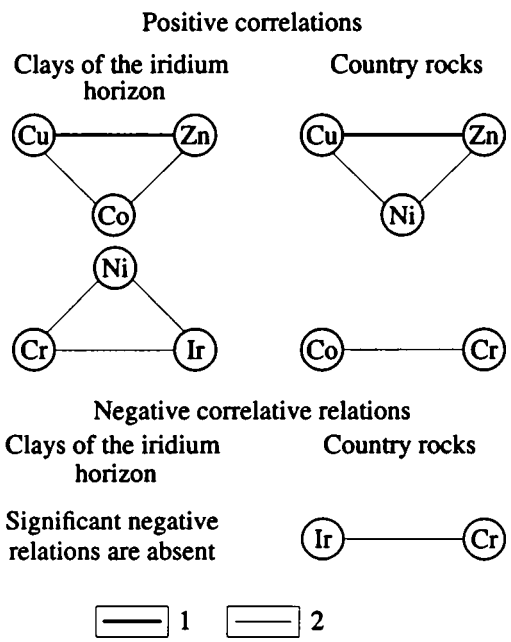

Fig. 1. *Schematic correlations between chemical elements in the iridium-horizon clays and the country rocks.*
1 – correlations with the significance level of 0.05 - 0.02;
2 – correlative relations with 0.1 - 0.05 significance level.

Table 3. Results of the factor analysis of the iridium-horizon rocks

Elements	Factors		
	I	II	III
Co	0.55	−0.38	−0.74
Cr	0.82	0.42	0.20
Cu	0.37	−0.84	0.22
Ni	0.87	0.32	0.08
Zn	0.57	−0.70	0.24
Ir	0.63	0.48	−0.08
Dispersions	43.05	30.96	11.66

Table 4. Results of the factor analysis of the country rocks of the iridium horizon

Elements	Factors		
	I	II	III
Co	0.27	0.76	0.45
Cr	−0.05	0.89	0.09
Cu	0.93	0.02	0.20
Ni	0.68	−0.02	−0.45
Zn	0.88	−0.18	0.00
Ir	0.01	−0.65	0.67
Dispersions	36.04	30.34	15.13

horizon makes it expedient to use indirect methods, such as geochemical comparisons of the rocks from the iridium horizon with the most probable sources of iridium. We believe this approach to be most fruitful.

An overwhelming majority of researchers believe that the chemical composition of iridium horizon rocks indicates that cosmic bodies (large meteorites, asteroids, or comets) or basaltic melts from the mantle could be the sources of substance for their formation. Therefore, we analyzed and compared the compositions of basic magmatic rocks (basalts and gabbroids) and stone meteorites (chondrites), which are the most representative variety of cosmic bodies. The data were accumulated from numerous publications, and 75 analyses formed massifs for each sample (magmatic rocks and meteorites). The concentrations of such elements as Cr, Co, Ni, Sc, La, Sm, Eu, Yb, and Lu marked the most substantial differences between the meteorites and the basalts. A comparison of the average contents of these elements, with the help of the Student t test and their dispersions (F-ratio), whose values considerably exceed the standards at a given significance level of 0.01, shows substantial geochemical and, apparently, genetic differences among the objects.

The next task is to carry out a geochemical comparison of the iridium horizon with one of the groups of rocks considered above. We used the elements according to whose distribution meteorites and basites are the most contrasting: Cr, Co, Ni, Sc, La, Sm, Eu, Yb, and Lu.

The discriminant function ($D = -0.034Sc + 0.023Cr - 0.048Co + 0.006Ni + 0.119\,La - 0.227\,Sm - 1.065\,Eu + 1.827\,Yb - 7.580\,Lu$; value $D_0 = 69.86$) and multiplicative ($KK_{mult} = \frac{Cr \cdot Co \cdot Ni \cdot Sc}{La \cdot Sm \cdot Eu \cdot Yb \cdot Lu}$) and additive ($KK_{add} = \frac{Cr + Co + Ni + Sc}{La + Sm + Eu + Yb + Lu}$) indices were used as the geochemical criteria. We analyzed 18 samples from the iridium horizon: 17 from three Turkmenistan sections (Alekseev *et al.*, 1988) and one from a Denmark section (Alwarez *et al.*, 1980). The calculations D, KK_{mult}, and KK_{add} are given in Table 5 and in Fig. 2. They testify unambiguously to the geochemical resemblance of the iridium horizon and basic magmatic rocks.

The results of our geochemical investigations are, perhaps, the only indisputable argument that proves the terrestrial, volcanic source of iridium in the formation of the horizon. Most likely, the biotic crisis that was

Table 5. Values of discriminant function, multiplicative and additive indexes for meteorites, basites, and iridium-horizon clays

Investigated objects and a number of observations	Geochemical indices		
	discriminant function	multiplicative	additive
Meteorites (75)	$\frac{133.06}{118.6 - 147.9}$	$\frac{1.06 \times 10^{16}}{1.7 \times 10^{15} - 1.4 \times 10^{17}}$	$\frac{2.1 \times 10^{14}}{1.3 \times 10^{4} - 2.9 \times 10^{4}}$
Basites (75)	$\frac{6.65}{-3.2 - 80.98}$	$\frac{1.46 \times 10^{13}}{7.3 - 9.3 \times 10^{14}}$	$\frac{454}{0.6 - 6.5 \times 10^{3}}$
Iridium-horizon clays	$\frac{3.27}{0.81 - 14.4}$	$\frac{4.2 \times 10^{6}}{5.7 \times 10^{4} - 3.6 \times 10^{7}}$	$\frac{14.7}{3.5 - 61.9}$

Note: The average values of geochemical indices are in the numerator; the limits of their variation values in the denominator.

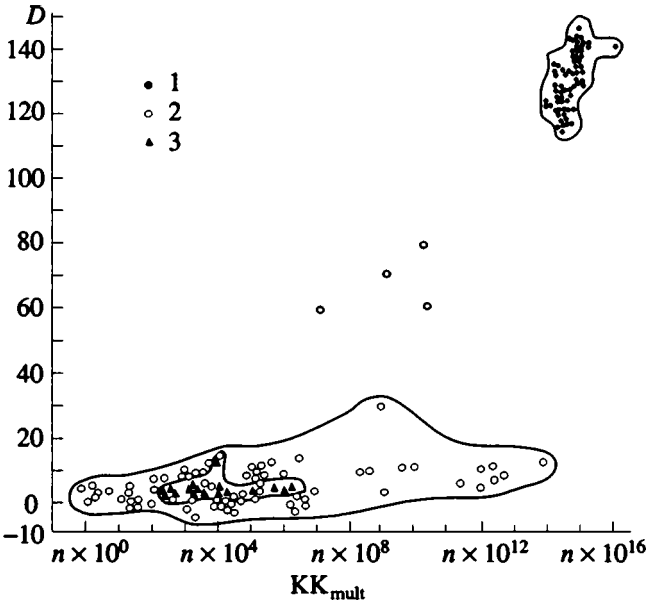

Fig. 2. *Correlation of D and KK_{mult} values in the iridium-horizon clays, basites, and meteorites.*
1 – meteorites; 2 – basalts; 3 – iridium-horizon clays; 4 – KK_{mult}.

clearly manifested in the "great" mass extinction and registed at the Cretaceous-Paleogene boundary at all taxonomical levels ought to be considered from these positions.

REFERENCES

Alekseev, A.S., Nazarov, M.A., Barsukova, L.D., *et al.*, The Cretaceous-Paleogene Boundary in the South of Turkmenistan and Its Geochemical Character, *Byull. MOIP, Otd. Geol.*, 1988, vol. 63, no. 2, pp. 55 - 69.

Alwarez, L.W., Alwarez, W., Asaro, F., and Michel, H.V., Extraterrestrial Cause for the Cretaceous-Tertiary Boundary Extinction, *Science*, 1980, vol. 208, no. 4448, pp. 1095 - 1108.

Brooks, R.R., Strong, C.P., Lee, J., *et al.*, Stratigraphic Occurrences of Iridium Anomalies at Four Cretaceous-Tertiary Boundary Sites in New Zealand, *Geology*, 1986, vol. 14, pp. 727 - 729.

Courtillot, V.E., A Volcanic Eruption, *Sci. Amer.*, 1990, vol. 263, no. 4, pp. 39 - 47.

Davis, J.C., *Statistics and Data Analysis in Geology*, John Wiley and Sons, New York, Brisbane, Toronto, Singapore, 1975.

Krumbein, W.C. and Graybill, F.A., *An Introduction to Statistical Models in Geology*, McGraw-Hill, New York, St. Louis, San Francisco, Toronto, London, Sydney, 1965.

Naidin, D.P., The Cretaceous-Paleogene Boundary at Mangyshlak and Suggested Events at the Maestrichtian-Danian Boundary, *Izv. Vyssh. Uchebn. Zaved., Geol. Razved.*, 1986, no. 9, pp. 3 - 12.

Naidin, D.P., Kopaevich, L.F., Moskvin, M.M., *et al.*, Micropaleontological Characteristic of the Maestrichtian and Danian in the Mangyshlak Continuous Sections, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1990, no. 12, pp. 68 - 82.

Nazarov, M.A., Badyukov, D.D., Barsukova, L.D., *et al.*, Parameters of the Cretaceous-Paleogene Shock Event, *Byull. MOIP, Otd. Geol.*, 1988, vol. 63, no. 4, pp. 33 - 54.

Neruchev, S.G., Global Geochemical Anomalies and Biosphere Crises, *Priroda*, 1988, no. 1, pp. 72 - 81.

Ronov, A.B., Khain, V.E., and Balukhovskii, A.N., *Atlas Litologo-Paleogeograficheskikh Kart Mira (Mezozoi i Kainozoi Kontinentov i Okeanov)* (Atlas of Lithologic-Paleogeographic Maps of the World (the Mesozoic and Cenozoic of Continents and Oceans)), Nedra, Leningrad, 1989.

Zakharov, V.A. and Lapukhov, A.S., Iridium Anomalies: Do They Always Testify to Impact Rearrangement of the Earth's Biosphere?, *37 Sessiya Vsesoyuzn. Paleontol. Obshch.* (37th Session of the USSR Paleontological Society), Nedra, Leningrad, 1991, pp. 33 - 34.

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Evolution of the biosphere

BRIEF COMMUNICATIONS

Climatic Rhythms in the Pleistocene of the East European Platform

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Abstract – Climatic rhythms were analyzed, which are represented by an alternation of oxygen-isotope stages in Pleistocene glacial and subaerial deposits in the central regions of the East European Platform. These correlate with transgressions of the Black Sea and with deposits of Belarus', the Baltic region, Poland, Germany, and the Netherlands.

As long ago as in the beginning of the 20th century, it was revealed that one of the characteristic features of the Pleistocene is an alternation of cold (glacial) and warm (interglacial) epochs. This inference, made first on the basis of studying the traces of the alpine glaciers onset, was successfully confirmed when an alternation of glacial and interglacial deposits was found in continental glaciation areas as well. These deposits were denoted by alpine terms: the Mindel, Mindel-Riss, Riss, Riss-Wurm, and Wurm. Soon it became clear, however, that the history of climatic fluctuations in the Pleistocene was by far more complex; the number of glacial and interglacial epochs singled out by different researchers was growing, which created certain problems with their correlation. Thus far, it has been difficult to overcome those problems despite involving, when correlating, the alternation of oxygen-isotope stages in oceanic sediments, loess and buried soils in subaerial deposits, and traces of marine transgressions and regressions. Therefore, the insights presented below are based on one of the plausible versions of correlation, which, from my point of view, is more substantiated, and which was used when compiling the regional stratigraphic scheme of the Quaternary for the central regions of the East European platform. The scheme was approved by the International Stratigraphic Commission in 1984 (Reshenie ... , 1986).

EARLY PLEISTOCENE

For the very beginning of the Pleistocene, all investigators single out a warm Mikhailovka (Petropavlovka) time, referred to the Matuyama epoch. The pertinent alluvial deposits comprise a peculiar assemblage of small mammals, transitional from the Taman' to Tiraspol type; malacofauna witnesses a warmer climate than that of the present. Well-expressed Balashev buried soil was formed on alluvial deposits. The position of the Brunes-Matuyama epoch boundary in the upper part of the Petropavlovka Horizon correlates with oxygen-isotope stages 19 and 20 (Fig. 1) and, accordingly, with the interglacial epoch I of the Cromerian in the Netherlands (Zagwijn, 1985).

The first Pleistocene cooling, corresponding to the Pokrovka Horizon, is well fixed in alluvial deposits by periglacial palynospectra and cryophilic mollusks and in subaerial deposits by a loess horizon. Possibly, it corresponds to the glaciation advanced until Moscow, though not all the investigators agree with a glacial origin of loamy soils of the Likhov sequence, which comprise pebbles and rubble of sedimentary rocks only. Pokrovka time can be correlated with the lower part of oxygen-isotope stage 18 and with glacial A of the Cromerian in the Netherlands.

The Il'inka Horizon accumulation coincided with the repeated climate fluctuations (Krasnenkov, 1990) that resulted in the formation of three fossil soils and three alluvial complexes (Fig. 1) in the upper Don River basin. These comprise the early Tiraspol' thermophilic mollusks and small mammals; the lowest complex also contains tortoise-shells. In the Moscow region, two interglacial units, separated by the Setun' moraine (Maudina *et al.*, 1985; Fursikova *et al.*, 1991), can, apparently, be attributed to Il'inka time. The upper Sukromnin unit correlates, by carpological data, with upper Il'inka (Moiseevka) layers, whereas the lower (Matveevka) one is most likely related to lower Il'inka beds. The Setun' moraine is well traced until the northern part of the Tula region, where its thickness decreases up to 0.5 - 1 m, wedging out near Yasnogorsk. In the Don River basin, the pebbles of crystal-

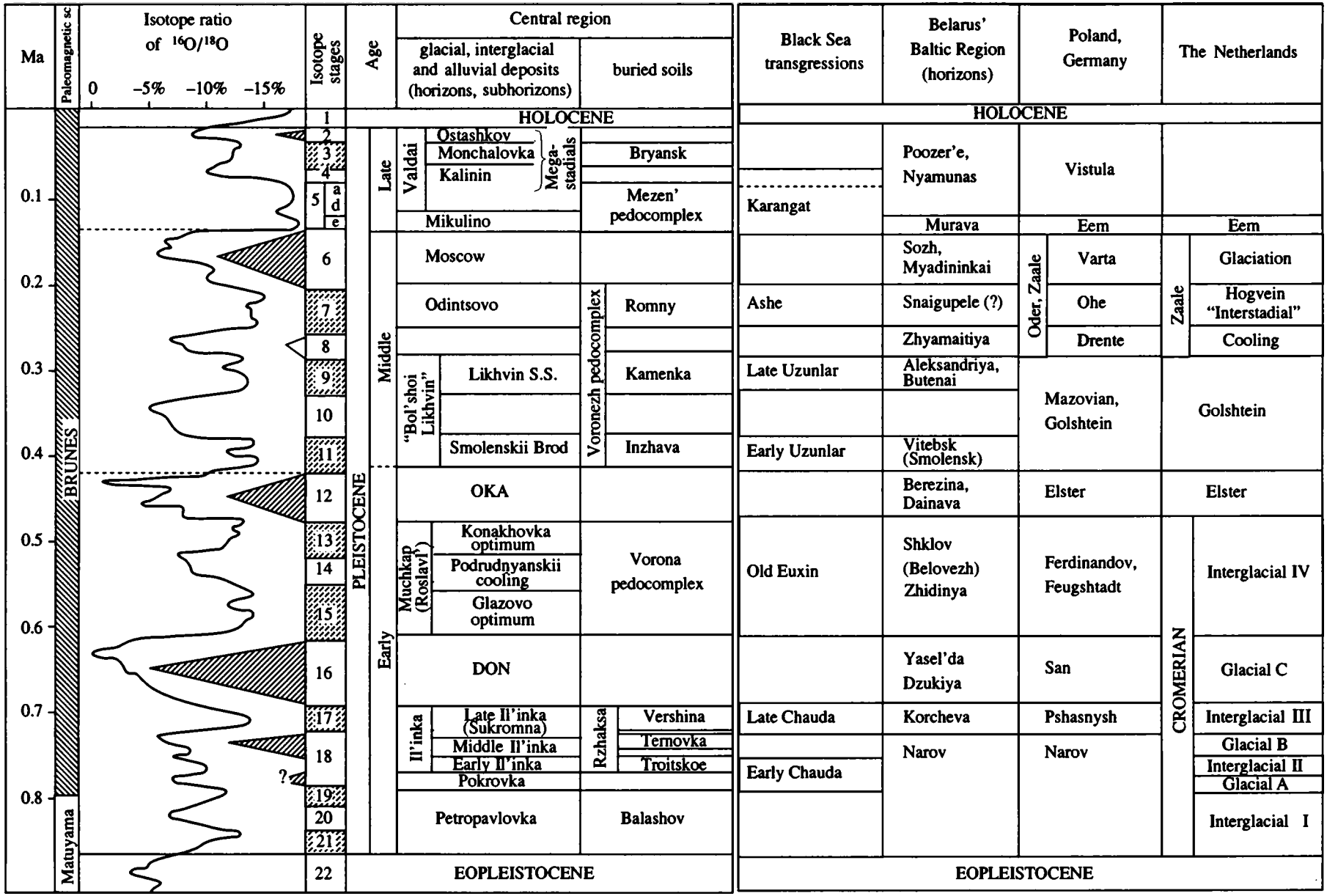

Fig. 1. *Correlation of the glacial and interglacial horizons with the oxygen-isotope curve. The hatching shows the relative dimensions of glaciers.*

line rocks, occurring at the base of upper Il'inka (Moiseevka) layers, correspond to this moraine. The upper Il'inka warm period can be correlated with oxygen-isotope stage 17, while the Setun' glaciation, with the strongest cooling at the end of stage 18, and the middle and early Il'inka warm episodes, with two warmings in the middle part of stage 18. In the Cromerian of the Netherlands, Interglacial II, Glacial B, and Interglacial III (Fig. 1) correlate with the Il'inka Horizon. Most likely, the middle Il'inka (Setun') moraine corresponds to the Narov moraine in Belarus' (though other schemes of correlation are feasible as well), and the interglacial epochs of Il'inka time, to the early and late Chauda transgression.[1]

The Don Glaciation, the most extensive in the early Pleistocene, corresponds to the coldest oxygen-isotope stage 16. Its moraine, producing the Don glacial tongue, is also welltraced in the Moscow region, Belarus', the Baltic region, and Poland, where it is distinguished under the names of Oster (Pereksha), Yasel'da, Dzukiya, and San, respectively; in the Netherlands, it correlates with the glacial C of the Cromerian.

The Don glaciation was followed by the very warm and long Roslavl' (Pol'nolapinsk, Muchkap) interglacial epoch, with two climatic optima separated by a strong cooling (and most likely, by a glaciation). The deposits of this interglacial epoch are characterized by the late Tiraspol' assemblage of small mammals. This epoch seems to encompass isotopic stages 15, 14, and 13, being well observed in Belarus' (Shklov), the Baltic region (Zhidinya), Poland (Ferdinandov), and Germany (Feugshtadt); in all places, the existence of two optima is typical for the epoch. In the Netherlands, this interglacial correlates with the interglacial of the Cromerian IV; in subaerial deposits of the southern part of the East European platform, with the very thick Vorona pedocomplex, and in the Black Sea region, with Old Euxin transgression.

The early Pleistocene terminates with the Oka Glaciation (isotopic stage 12), which covered the entire territory of the Smolensk region and only the northern part of the Moscow region. Its moraine is well traced in Belarus' (the Berezina moraine), the Baltic region (the Dainava moraine), Poland (the Vilga moraine), and West Europe (the Elster moraine).

MIDDLE PLEISTOCENE

The middle Pleistocene starts with a far more complicated Likhvin interglacial epoch: the "Bol'shoi Likhvin." The data available promotes distinguishing two climatic optima in it (though in the central regions, these are not revealed in the same section). The earliest optimum is represented by deposits of the Smolenski Brod exposure near Velizh, where the oldest species of the genus *Arvicola* were found; the latest optimum (the Likhvin proper), by deposits of the stratotype section near Chekalin, with more evolved representatives of this genus (Voznyachuk and San'ko, 1981; Motuzko, 1984). Glacial deposits, separating these optima, are observed nowhere; if glaciation occurred at this time, it would cover a very small area. The Bol'shoi Likhvin interglacial epoch seems to encompass the isotopic stages 11, 10, and 9; its deposits are reliably traced in Belarus', in the Baltic region, Poland, and West Europe (the Aleksandriya, Butenai, Mazovets, and Goldshtein interglacials, respectively, where sometimes two optima are observed). In subaerial deposits, the Bol'shoi Likhvin interval correlates with two lower buried soils (the Inzhava-Kamenka) of the middle Pleistocene loess-soil complex, while in the Black Sea region, it correlates with the early and late Uzunlar transgression.

Different opinions exist on the limits of glaciation following the Likhvin interglacial. Many investigators attribute the moraine of the Dnieper glacial tongue to it. Recently, however, data appeared supporting the moraine formation at the end of the middle Pleistocene, rather than in its first half. In the southern parts of the Bryansk region, the Mikulino interglacial deposits were recorded in residual hollows on the surface of the moraine stretching southward to the Ukraine, and for the moraine proper as well as for the overlying and underlying fluvio-glacial deposits, a series of radiometric dates, from 146 ka to 170 ka, were obtained (determinations of O.A. Kulikov). Apparently, the post-Likhvin Glaciation (isotopic stage 8) did not reach the central regions because its moraine is seen only in the Baltic region, where it is distinguished as the Zhyamaitiya moraine. The idea that the strongest glaciation was in the late middle Pleistocene correlates well with data obtained by Zagvein (1985), who singled out the early Zaale cooling and late Zaale Glaciation.

The question on the middle Pleistocene post-Likhvin interglacial epoch, corresponding to isotopic stage 7, is debatable as well. In the central regions, the interglacial character of this stratigraphic interval was argued by A.I. Moskvitin (1946), who considered the buried soil, observed below the Moscow moraine in a quarry near Odintsovo, as its stratotype. Recently, M.I. Maudina (personal communication) discovered that this buried soil was not replaced laterally by interglacial deposits, as V.N. Sukachev *et al.* (1969) supposed, but it overlies the lacustrine series of complex structures, where two optima of the Roslavl' interglacial epoch are distinguished, as well as the Likhvin deposits, or deposits accumulated as a result of their reworking (Breslav *et al.*, 1979). Thus, this confirms the post-Likhvin age of the buried soil in the Odintsovo section, which retains the name Odintsovo for the post-Likhvin interglacial epoch. However, to this day, nowhere in the central regions were found the lacustrine-swampy deposits of this age. In the Baltic region, the Snaigupele deposits

[1] Chronology of the Black Sea transgressions is given after Ya.I. Izmailov (*Presentation at All-Union Meeting on the Quaternary*, Tallinn, 1990), while stratigraphy of subaerial deposits, after A.A. Velichko *et al.* (1984); for Belarus', the materials of F.Yu. Velichkevich *et al.* (1992) were used, and for Poland, of I.E. Mojski (1985).

seem to be synchronous with this stratigraphic interval, though F.Yu. Velichkevich (1982) considers the Snaigupele flora to be older than that of the Likhvin.

In West Europe, the Ober-Ohe and other sections of the Luneburger Heide locality (Selle, 1954; Hoverman *et al.*, 1985) are the most feasible analogs of the middle Pleistocene post-Likhvin interglacial deposits; in the south of the East European platform, the Romny fossil soil is the most feasible analog, and in the Black Sea region, they are the Asha transgression deposits.

The boundary of the advance of the Moscow Glaciation (isotopic stage 6), the last in the middle Pleistocene, is also still debatable. As it was previously shown, it was the strongest glaciation in the middle Pleistocene, being related to the formation of the Dnieper glacial tongue, lacking the glacial deposits older than the middle Pleistocene.[2]

LATE PLEISTOCENE

Deposits of the Mikulino interglaciation, which started in the late Pleistocene, are studied in detail in many sections, and they correlate well everywhere, both in East and West Europe (the Murava strata in Belarus', Myarkino in the Baltic region, and Eemian strata in western regions). In subaerial deposits, they undoubtedly correspond to the lower part of the Mezen' pedocomplex, and in the Black Sea region, they correspond to the Karangat transgression strata. However, not all the researchers are unanimous when correlating the Mikulino interglaciation with the oxygen-isotope scale; most likely, it corresponds only to the lower part of stage 5 (phase 5e).

The chronology of climatic events in Valdai time causes no contradictions. Two cold (glacial) epochs are singled out in it: the early Valdai epoch (the Kalinin), corresponding to isotopic stage 4 and to the upper part of stage 5, and the late Valdai epoch (the Ostashkov), which can be correlated to stage 2. Most of the researchers consider a long epoch (about 25 ka) separating them and corresponding to isotopic stage 3, as a megainterstadial, with the climate, in general, cooler than that of the present-day, though repeated alternation of the warmings and coolings occurred at that time. Such a concept is consistent with the oxygen-isotope data, according to which the corresponding warming on the paleotemperature curve shows a lower amplitude in comparison with the warmings correlating with the interstadials. The formation of the Bryansk buried soil appears to coincide with this time.

The boundary of the late Valdai (Ostashkov) Glaciation is traced very clearly and does not add any contradictions. The early Valdai Glaciation boundary is more complicated. Some investigators believe it to be stronger than the late Valdai Glaciation and to stretch until the northern Moscow region. The observations available show that south of the Ostashkov Glaciation boundary, the Mikulino deposits are not overlain by the moraine; only at the edges of hollows, to which those deposits are confined, do the moraine-like deposits of deluvial-solifluction origin occur, being absent in their central part. Apparently, the early Valdai Glaciation was considerably weaker than the late Valdai Glaciation.

Thus, at least four glaciations are distinguished in the early Pleistocene; the Don Glaciation, the largest of them, correlates with its middle part and has been the strongest throughout the entire Pleistocene (Fig. 1). In the middle Pleistocene, the last of the two or three most recent glaciations was the strongest; in the late Pleistocene, the last glaciation was maximum as well.

The history of Pleistocene glaciations appears to be more complicated, as compared with Yu.A. Lavrushin's concept (1991), who believed the Mindel-Riss to divide the glacial Pleistocene into two long glacial epochs: the pre- and post-Mindel-Riss stages. First, the Muchkap stage, which Lavrushin included into the Don glacial epoch, seems to be on par with the Mindel-Riss stage, in terms of both duration and climatic conditions; secondly, following the Mindel-Riss stage, no glaciations occurred.

[2] This glaciation should be called the Dnieper glaciation, but it might cause confusion, as many investigators consider the Moscow and Dnieper moraines to be diachronous.

REFERENCES

Breslav, S.L., Valueva, M.N., and Maudina, M.I., New Data on the Odintsovo Stratotype Section, *Dokl. Akad. Nauk SSSR*, 1979, vol. 248, no. 1, pp. 161 - 166.

Fursikova, I.V., Pisareva, V.V., Yakubovskaya, T.V., *et al.*, The Startotypical Section of the Pleistocene at Okatovo Village in the Western Part of the Moscow Region, *Stratigrafiya Fanerozoya Tsentra Vostochno-Evropeiskoi Platformy*, Tsentrgeologiya, Moscow, 1992, pp. 59 - 82.

Khoverman, Yu., Chebotareva, N., and Pisareva, V., On the Age of Kieselguhrs at Ohe Settlement (Luneburger Heide, Germany), *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1985, no. 4, pp. 39 - 48.

Krasnenkov, R.V., Stratigraphy and Geochronology of the Lower Pleistocene of Eastern and Central Europe, *Chetvertichnyi Period: Metody Issledovaniya, Stratigrafiya i Ekologiya. VII Vsesoyuzn. Soveshch.*, Akad. Nauk Estonii, Tallinn, 1990, vol. 2, pp. 91 - 92.

Lavrushin, Yu.A., Types of Geological and Paleoecological Environments in the Quaternary, *Geologo-Paleoekologicheskie Obstanovki Chetvertichnogo Perioda (k XIII Kongressu INKVA)*, Geol. Inst. Akad. Nauk SSSR, Moscow, 1991, pp. 7 - 17.

Maudina, M.I., Pisareva, V.V., and Velichkevich, F.Yu., The Odintsovo Stratotype Section in the Light of New Data, *Dokl. Akad. Nauk SSSR*, 1985, vol. 284, no. 5, pp. 1195 - 1199.

Mojski, J.E., Geology of Poland, *V., Stratigraphy. Part 38. Cainozoic. Quaternary*, Wyd. Geol., Warshava, 1985.

Moskvitin, A.I., The Odintsovo Interglacial, and Position of the Moscow Glaciation among Other Glaciations of Europe, *Byull. MOIP, Otd. Geol.*, 1946, vol. 21, no. 4.

Motuzko, A.N., Anthropogene Rodents in Belarus' and Adjacent Areas, *Problemy Pleistotsena*, Nauka i Tekhnika, Minsk, 1984, pp. 173 - 188.

Reshenie 2-go Mezhvedomstvennogo Stratigraficheskogo Soveshchaniya po Chetvertichnoi Sisteme Vostochno-Evropeiskoi Platformy (Leningrad–Poltava–Moscow, 1983) (Resolution of the 2nd Inter-Departmental Stratigraphic Conference on Quaternary System of the East European Platform), VSEGEI, Leningrad, 1986.

Selle, W., Die Vegetationsentwicklung des Intergluzials von Ober-Ohe in der Luneburger Heide, *Abh. Naturw.*, Verein, Bremen, 1954, vol. 33, no. 3, pp. 457 - 469.

Sukachev, V.N., Gorlova, R.N., and Nedoseeva, A.K., On the Pleistocene Flora at Odintsovo Station near Moscow, *Byull. Komissii po Izucheniyu Chetvertichnogo Perioda Akad. Nauk SSSR*, 1969, no. 36, pp. 3 - 7.

Velichkevich, F.Yu., *Pleistotsenovye Flory Lednikovykh Oblastei Vostochno-Evropeiskoi Ravniny* (The Pleistocene Floras of the Glacier-Covered Areas of the East European Plain), Nauka i Tekhnika, Minsk, 1982.

Velichko, A.A., Markova, A.K., Morozova, T.D., and Udartsev V.P., Problems of Geochronology and Correlation of Loess and Fossil Soils of Eastern Europe, *Izv. Akad. Nauk SSSR, Ser. Geogr.*, 1984, no. 6, pp. 5 - 19.

Voznyachuk, L.N. and San'ko, A.F., The Stratotype Section of the Pleistocene in Smolensky Brod Locality, *Kompleksnoe Izuchenie Opornykh Razrezov Nizhnego i Srednego Pleistotsena Evropeiskoi Chasti SSSR, Geol. Fond RSFSR*, Moscow, 1981, pp. 50 - 54.

Zagwijn, W.H., An Outline of the Quaternary Stratigraphy of the Netherlands, *Geologie Mijnbouw*, 1985, vol. 64, pp. 17 - 24.

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BRIEF

COMMUNICATIONS

Upper Paleozoic of the South Emba Gravity Maximum Area

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At the end of 1930s, northeast of the Buzachi Peninsula, by means of pendulum and gravimetric survey, several large positive gravity anomalies were revealed and named as the South Emba gravity maximum. At the end of 1940s, deep drilling was started on the cratonic structures set by seismic survey according to Mezozoic datums integrated into the South Emba uplift. For the upper Paleozoic deposits, it was revealed in deep boreholes 4,7-Turesai and 10,11-Zhanasu, which exposed the Lower Carboniferous and Upper Devonian terrigenous sediments under the Jurassic layers (Aizenshtadt *et al.*, 1957; Grachev, 1959; Garetskii *et al.*, 1961; Dneprov, 1962; Yanshin *et al.*, 1961). The data obtained was widely used and is still in considerable use in the interpretation of the seismic data (Voronov and Nazhmetdinov, 1979; Nazhmetdinov *et al.*, 1979; Turkov and Umirshin 1980; Umirshin, 1983) and for the estimation of oil-and-gas-bearing prospects (Turkov, 1980; Umrishin, 1983). This served to increase the amount of geological exploration in the area. However, boreholes, already deeply drilled, were inefficient, providing a reason for reducing the prospecting and exploratory drilling, which, in our opinion, was unjustified. Such a decision was due to the inadequate study on the stratigraphy of the sections penetrated by the boreholes, though many investigators paid proper attention to that question (Benenson *et al.*, 1978; Sinel'nikov, 1987).

On the Russian platform and Caspian depression, including its eastern part, the Carboniferous and Upper Devonian deposits are composed largely of carbonates (Aizenshtadt *et al.*, 1957; Zolotukhina *et al.*, 1986, 1988, 1989). Boreholes 4,7-Turesai and 10,11-Zhanasu, however, penetrated into terrigenous strata, which some authors (Garetskii *et al.*, 1961) correlated with the Zilair group of the south Urals as having no analogs within the boundaries of ancient cratons.

During the last 10 years, west and northwest from Turesai and Zhanasu (in Tengiz, Yuzhnaya, Korolevskaya, Tazhigali, Saz-Tube, Arman, Pionerskaya, and other areas), a number of boreholes were drilled, penetrating into Devonian and Carboniferous sediments in carbonate facies overlain by a carbonate-terrigenous unit of late Artinskian age (Zolotukhina *et al.*, 1986, 1988, 1989). In close proximity to the southern termination of the South Emba maximum, boreholes 1-Saz-Tube, 1,2-Eastern Arman, and 2-Arman revealed a carbonate-terrigenous unit under Triassic deposits (drilled to 1288, 1093, 1011, and 622 m, respectively). The unit is composed of granule gravelstone, conglomerate, claystone, siltstone, sandstone, and breccia-conglomerate. The latter consists of rather large limestone fragments of different textures. The unit was subdivided by O.B. Ketat (Zolotukhina *et al.*, 1988) into five members and referred to the rudaceous type of the late Artinskian sequences.

Member I, drilled to more than 200 m, is composed of claystone and polymictic siltstone, with an admixture of pyroclastic material and interbeds of granule gravelstone and breccia-conglomerate, largely of pyroclastic composition, containing rare fragments of calcarenite and mud-detrital limestone with the silicified faunal and floral remnants *Archaediscus* sp. and *Koninckopora* sp., redeposited from Carboniferous sediments (borehole 2-Eastern Arman, depth of 4805 - 4087 m).

Member II is represented by siltstone, sandstone, claystone, and calcareous breccia-conglomerate, with a considerable content of extrusive material.

The breccia-conglomerate is composed of limestone fragments of different textures: grumous, mud-detrital, and fine-grained. In the detrital part, the following heterochronous organic remains were found: (1) Frasnian-Famennian and Carboniferous *Rectangulina* sp., *Asterosphaera pulchra* Reitl., *Archaediscus* sp., *and Mediocris ovalis cupellaeformis* (Gan.), (borehole 2-Pionerskaya, depth 4854 - 4878 m); (2) Early and Middle Carboniferous and Asselian *Howchinia gibba* Noell., *Cornospira* sp., *Ozawainella nikitovkensis* (Brazh.), and *Pseudofuzulina* sp., conodonts *Streptognathodus constrictus* Chern. and *Neogonolella* sp. (determinations by N.S. Ovnatanova) (borehole 5-Yubileinaya, depth of 4803 - 4810 m).

The cement of the breccia-conglomerate is represented by clayey-bituminous material with numerous sponge spicules and radiolarian remains forming spongolite and radiolarite interbeds.

The claystone and spongolite of member II in the section, of borehole 10-Munaibai (depth 5235 - 5238 m), yield numerous *Tasmanites* algae (sometimes amounting up to 25% of the rock in thin sections) and microfossils in which late Artinskian *Vittatina* (35%), *Entylissa* (15%), *Monosaccites* (15%), *Striatiti* (15%), and *Astriatiti* (15%) are dominant. Subdominant species are represented by *Vittatina vitifer* Lub., *V. subsaccata* Lub.,

V. costabilis (Lub.) Lub., and others (microfossil determinations here and after by E.N. Zdobnova).

Member III, 122 - 204 m thick, is composed of calcareous granule gravelstone and breccia-conglomerate with spongolite and claystone interbeds. In spongolites, the small, late Artinskian foraminifers *Nodosaria shikanica* Lip. and *Geinitzina minima* Raus. were found, along with the microfossils *Vittatina costabilis* Wil., *Calamospora hartungiana* Schopf., and *Acanthozonotriletes connutus* Schwartz. In the detrital part, strongly silicified heterochronous Middle Carboniferous to Early Permian foraminifers and algae are abundant.

Member IV, 135 - 245 m thick, is composed of claystone and micrograined limestone with spongolite, radiolarite, and thin lenticular silicified calcareous granule gravelstone interbeds. In the latter, small sediments, redeposited from the lower Bashkirian, with the foraminifers *Archaediscus* and *Asteroarchaediscus* were found. In addition, in micrograined limestone, late Artinskian *Nodosaria mirabilis* Lip., *Geinitzina minima* Raus., and *Heperamninoides samarensis* Raus. and *Tasmanite algae* were found (borehole 2-Yubileinaya, depth 4428 - 4403 m; borehole Saz-Tube, depth 3817 - 3823 m).

The sediments of member V, 45 - 100 m thick, are represented by marly limestone (Tengiz area) or calcareous and silty claystone, with thin interbeds of granule gravelstone and breccia-conglomerate. The latter (in clastic part) is composed of fragments and intact specimens of Middle Carboniferous, Asselian, and Sakmarian fusulinids. In the cement, late Artinskian *Geinitzina minima* Raus. and *Nodosaria longissima* Sul., radiolarians, and sponge spicules are noticed (borehole 1-Pionerskaya, depth 4228 - 4234 m).

While correlating the carbonate-terrigenous strata of the southeastern part of the Caspian basin from the west eastward, from Tengiz to Eastern Arman, we found certain regularities in their structure:

(1) Changes in the lithologic characteristics of members due to their position relative to the source;

(2) Variations both in size, roundness, and sorting of grains and in preservation, abundance and age of redeposited flora and fauna in sections;

(3) Similar structure of upper Artinskian sediments, usually represented by relatively deep-water radiolarites, spongolites, and micrograined limestone, yielding impoverished euryfacies assemblages of small foraminifers ecologically incompatible with earlier benthic stenofacies fusulinids.

The age of the terrigenous unit of the South Emba uplift lacks sufficient faunal substantiation.

From the lower part of the South Emba formation, a spore-and-pollen assemblage of both the Early and Middle Carboniferous was recovered (Grachev, 1959). Nevertheless, it was ignored, and sediments yielding that mixed assemblage were referred to the middle Visean (Bobriki Substage). Yanshin *et al.*, (1961) decided to rectify this error, regarding the age of the South Emba formation as younger, but still Early Carboniferous. In the work by Benenson *et al.*, (1978) it is shown, with reference to the Turesai area, how the same unit was dated differently by different investigators.

According to common opinion, the upper part of the terrigenous unit in the Tortai area is represented by limestone. However, Sinel'nikov (1987) found it to be represented by rudaceous carbonate-terrigenous rocks composed of limestone fragments in which Late Carboniferous and Early Permian organic remains (most likely redeposited) were found and identified. Benenson *et al.*, (1978) reject the possibility of facies replacement of carbonate rocks by terrigenous ones between boreholes G-3 and G-4-Turesai and attribute the lack of the former in borehole G-4 to their erosion.

Lithologic and paleontologic data on new boreholes located in the immediate vicinity of the South Emba maximum, with regard to the former studies (Benenson *et al.*,1978; Sinel'nikov, 1987), makes it possible to consider the carbonate-terrigenous unit found in the Arman, Eastern Arman, Saz-Tube, and other areas to be identical to the terrigenous unit of the South Emba uplift.

To make the investigation of promising Lower Permian deposits more efficient, it is necessary to carry out an additional, comprehensive lithologic and petrographic analysis and a careful paleoecologic study of autochthonous and allochthonous fossils.

Therefore, the Carboniferous and Devonian carbonate deposits occurring in the South Emba uplift *in situ*, which proved to be highly productive (where terrigenous rocks are present), were not, most likely, yet penetrated by boreholes.

REFERENCES

Aizenshtadt, G.E., Lapina, N.N., and Stepanov, D.L., New Data on the Geology of the Southeastern Russian Craton, *Geol. Geokhim.*, Gostoptekhizdat, Leningrad, 1957, issue 7, pp. 58 - 75.

Benenson, V.A., Kunin, N.Ya., Morozova, M.N., and Nurzhanov, K.K., *Paleozoiskie Otlozheniya Pogranichnykh Raionov Turanskoi i Russkoi Plit* (Paleozoic Deposits of Bordering Areas of Turan and Russian Platforms), Nedra, Moscow, 1978.

Dneprov, V.S., *Geologicheskoe Stroenie i Neftegazonosnost' Yuzhno-Embinskogo Podnyatiya i Severnogo Ustyurta* (Geology and Oil-and-Gas Presence in the South Emba Uplift and Northern Ustyurt), Gostoptekhizdat, Leningrad, 1962.

Garetskii, R.G., Egorov, I.P., Naumova, S.N., and Shlezinger, A.E., Lower Carboniferous and Upper Devonian Deposits of the Zhanasu Area (South Emba Gravity Maximum), *Dokl. Akad. Nauk SSSR*, 1961, vol. 136, pp. 1418 - 1421.

Garetskii, R.G., Naumova, S.N., and Shlezinger, A.E., Stratigraphy and Facies Groups of Upper Devonian Deposits in the South Emba Gravity Maximum Area, *Dokl. Akad. Nauk SSSR*, 1961, vol. 141, no. 4, pp. 931 - 933.

Grachev, R.I., Geology and Evolution of South Emba Uplift, *Geol. Sb.*, Gostoptekhizdat, Leningrad, 1959, no. 4, pp. 215 - 265.

Nazhmetdinov, V.Sh., Umirshin S.K., and Voronov, G.V., Oil-and-Gas Prospects of Sediments Underlying the Salines in the Southeastern Caspian Basin, *Prognoz Neftegazonosnosti Kazakhstana i Sopredel'nykh Territorii, Trudy VNIGNI*, Moscow, 1979, issue 212, pp. 169 - 174.

Synel'nikov, A.V., On the Role of Minute Lithologic and Petrographic Studies in Elaboration of the Stratigraphic Base (With Reference to the Tortai Area), *Stratigrafiya i Paleontologiya Paleozoya Prikaspiiskoi Vpadiny*, VNIGNI, Moscow, 1987, pp. 114 - 121.

Turkov, O.S. and Umirshin, S.K., Geology and Oil-and-Gas Content in the Undersaline Tortai Deposit, *Geol. Nefti i Gaza*, 1980, no. 9, pp. 57 - 61.

Umirshin, S.K., Structure and Oil-and-Gas Prospects of the Southeastern Margin of the Caspian Basin, *Geologiya i Perspektivy Neftegazonosnosti Bortovykh Zon Prikaspiiskoi Vpadiny*, VNIGNI, Moscow, 1983, pp. 102 - 105.

Voronov, G.V. and Nazhmetdinov, A.Sh., Paleozoic Carbonate Sediments of the South Emba Uplift, *Neftegaz. Geol. i Geofiz.*, 1979, no. 7, pp. 15 - 17.

Yanshin, A.L., Garetskii, R.G., Naumova, S.N., and Shlezinger, A.E., On the Position of the Russian Craton Boundary East of the Caspian Sea, *Byull, MOIP, Otd. Geol.*, 1961, vol. 36, no. 4, pp. 76 - 96.

Zolotukhina, G.P., Ketat, O.B, Taboyakova, V.Ya., *et al.*, Lithologic and Stratigraphic Characteristics of Sediments Underlying the Salines in the Tengiz Area, *Geol. Nefti i Gaza*, 1986, no. 1, pp. 22 - 26.

Zolotukhina, G.P., Ketat, O.B, Dan'shina, N.V., and Aleshin, V.M., Stratigraphy and Accumulation Conditions of Paleozoic Sediments of the Tengiz Deposit, *Razvedka i Burenie Neftyanykh Mestorozhdenii*, VNIGNI, Moscow, 1988, pp. 17 - 26.

Zolotukhina, G.P., Ketat, O.B, Dan'shina, N.V., and Aleshin, V.M., On the Stratigraphy of Upper Paleozoic deposits in the Southeastern Caspian Basin (Tengiz Area), *Dokl. Akad. Nauk SSSR*, 1988, vol. 304, no. 4, pp. 933 - 936.

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